Univerzita Karlova

Přírodovědecká fakulta

Studijní program: Geologie

Studijní obor: Geologie

Dominika Linzerová

Mechanismy vzniku kumulátů v mafických intruzích

Processes of cumulate formation in mafic intrusions

Bakalářská práce

Vedoucí práce: Mgr. Václav Špillar, Ph.D.
Konzultant práce: doc. Alessandro Fabbrizio, Ph.D.

Praha, 2021

Prohlášení / Statement of authorship

I declare that this work, completed under the supervision of Mgr. Václav Špillar, Ph.D., is a result of my own research and that I have cited all used resources and literature. Neither this thesis nor its substantial part has been submitted to fulfil requirement for other or coequal academic degree.

Prohlašuji, že jsem závěrečnou práci zpracovala samostatně pod vedením Mgr. Václava Špillara, Ph.D. a že jsem uvedla všechny použité informační zdroje a literaturu. Tato práce ani její podstatná část nebyla předložena k získání jiného nebo stejného akademického titulu.

V Praze / In Prague, 24.8.2021 Dominika Linzerová

Acknowledgements

The thesis was written at the Institute of Petrology and Structural geology, Faculty of Science, Charles University, and all experiments were performed in the Laboratory of Experimental Petrology.

I wish to express my deepest gratitude to my supervisor Václav Špillar, who greatly inspired me and helped me overcome any difficulties encountered during completion of this work. Without his guidance and kind approach, the goals of this study would not have been reached. It was my greatest pleasure to begin my journey as a magmatic petrologist under his supervision, and I hope that this work is merely the first out of our many future accomplishments.

I would also like to pay my regards to Alessandro Fabbrizio for his helpful advice during the experimental work, and to Martin Racek for his assistance in the SEM laboratory.

Abstract

Igneous cumulates and their parental intrusions are best known amongst geologist for their economical profitability as the main source of platinum group elements. However, beyond the scope of mining interests, cumulate rocks represent one of the most complicated and least understood topics in igneous petrology. Their formation is assumed to be driven by fractional crystallization of mostly mafic parental magmas, achieved by a separation of crystals from their surrounding liquid. In this work, I review the extend of current knowledge regarding the wide variety of possible formation processes, as well as the common properties and settings of cumulates and means of their classification.

In order to demonstrate the operation of the reviewed processes we carried out a series of high temperature experiments, investigating a behaviour of two different crystal suspensions over a period of 20 hours. Experiments were performed in a classic 1-atm. furnace at 1350 and 1390 °C, using a 10 wt. % and 60 wt. % mixtures of natural olivine and synthetic haplobasaltic glass. In a liquid rich environment, we observed a progressive formation of a cumulate layer by crystal settling, while in contrast, the suspension containing 60 wt. % of olivine effectively inhibited all movement aside from expansion of bubbles trapped inside the crystal mush. Textural analysis of all experimental samples revealed an effect of compositional convection, which enabled faster dissolution of small grains in the liquid-rich experiments, thus more rapid increase of the median grain size. Spatial distribution pattern of olivine crystals in both sample sets evolves from random to anti-clustered (i.e., ordered). This trend is more prominent in low crystallinity samples, where it is attributed to the effect of crystal redistribution. Our experiments provide qualitative record of mechanical and kinetic processes reported from the mafic cumulates worldwide.

Abstrakt / Czech abstract

Magmatické kumuláty a intruze které spoluutváří jsou mezi geology známé především jako ekonomicky významný zdroj platinových kovů, avšak stranou od jejich těžebního potenciálu představují kumulátové horniny jedno z nejvíce komplikovaných a nejméně pochopených témat magmatické petrologie. Jejich vznik je obecně spojován s frakční krystalizací především mafických magmat, která je řízená separací krystalů od taveniny v jejich okolí. Tato práce shrnuje základní prvky současného poznání o široké škále procesů zodpovědných za jejich formování, a zároveň popisuje typické znaky kumulátových hornin, prostředí jejich výskytu a způsoby klasifikace.

Za účelem demonstrovat fungování popsaných procesů jsme provedli sérii vysokoteplotních experimentů, zkoumajících průběh chování dvou různých krystalových suspenzí v časovém rámci dvaceti hodin. Všechny experimenty probíhaly v klasické peci za atmosférického tlaku a teplotách 1350 a 1390 °C, za použití směsí obsahujících 10 a 60 hm. % přírodního olivínu a syntetické bazaltové sklo. První ze suspenzí představující prostředí s nízkou počáteční krystalinitou poskytla možnost pozorovat průběh vzniku kumulátové vrstvy procesem sedimentace, zatímco směs obsahující vysoký podíl olivínových krystalů efektivně zabránila jakémukoli pohybu uvnitř kapsule s výjimkou nucené expanze bublin uvězněných v krystalové kaši. Texturní analýza vzorků odhalila vliv kompoziční konvekce, která umožnila urychlení rozpouštění drobnějších zrn v taveninou bohatém prostředí, a tudíž i rychlý vzrůst mediánu velikostí zrn. Prostorová distribuce olivínových zrn se v obou experimentální setech vyvíjí od náhodného až do uspořádaného rozmístění, přičemž je tento trend výraznější u vzorků s nižší počáteční krystalinitou kde je přisuzován efektu redistribuce zrn. Naše experimenty tak poskytují kvalitativní záznam mechanických a kinetických procesů, které jsou pozorovány v mafických kumulátech z celého světa.

Contents

I. Introduction

Igneous cumulates are texturally and compositionally variable group of rocks, characterized by a framework of crystals which formed and accumulated through fractional crystallization of their parental magma. Due to the nature of their formation, cumulates require low viscosity conditions and liquid-rich environment, hence are found predominantly, but not exclusively, in solidified basaltic magma chambers, occasionally sills and thicker lava flows (Irvine 1982, Holness 2020).

Every cumulate rock consists of two essential structural elements: The primary crystals making up the framework, known as primocrysts, and the later crystallizing interstitial phases in the pore spaces between them. A traditional cumulate nomenclature uses the amounts of these two basic components to separate the rocks into several categories, each characterized by modal range of interstitial phases. Therefore, the terminology lies on a scale with two extreme members: Orthocumulates, which contain more that 20 % of the interstitial material, and adcumulates, with as little as 5 % or less (Wager et al 1960, Wager 1963). It is the latter that causes the most friction in the field of cumulate research, as the processes leading to the reduction of framework porosity and formation of adcumulates are not fully understood (Holness et al 2017).

The basic principle behind cumulate formation is a separation of primocrysts from the surrounding liquid. The separation can be achieved in two different ways: Either by a redistribution of primocrysts from their initial location and accumulation elsewhere (Wager et al 1960), or by an expulsion of interstitial liquid while the crystals grow in-situ (McBirney and Noyes 1979, Boudreau 2011). Each of these contrasting mechanisms comprises a variety of possible processes driving the separation, all of which can operate throughout one magma chamber simultaneously (Sparks et al 1985). This results in a fascinating textural diversity amongst cumulates, but on the other hand severely complicates the efforts to decode their formation history (Hunter 1996).

Layered mafic intrusions represent the largest exposures of cumulate rocks on Earth, and thus have played a pivotal role in every fundamental cumulate study throughout almost seventy years of research (e.g., Wager and Deer 1939, Campbell 1978, McBirney and Noyes 1979, Boudreau 2011). Making up the signature layered sequences, the cumulates display a stunningly detailed record of their solidification history and are often taken as natural laboratories for observations and study of magma chamber processes in general (Holness et al 2017). Aside from their research potential, cumulate rocks in layered intrusions are greatly valued amongst geologists as an indispensable source for a variety of ores and noble metals, mainly the platinum group elements (Lee 1996).

In this work, I attempt to summarize the essentials of current knowledge regarding the cumulate rocks, with a primary focus on their formation mechanisms. The processes theoretically described on the

following pages are then investigated in a series of high-temperature experiments, using a mixture of olivine crystals and basaltic glass to simulate a suspension of primocrysts in a natural mafic intrusion. To make the experiments relevant for a wider range of conditions, two types of mixtures were used in order to produce both liquid-rich and crystal dominated environments.

II. Igneous cumulates

Igneous cumulates are a group of rocks formed by the separation of crystals from the main liquid in the early stages after the emplacement of the magma body. The separation is achieved either by redistribution of crystal from the place of their initial development and their accumulation elsewhere, or by in-situ nucleation and growth of crystals accompanied by separation of the liquid. Each cumulate consists of two main components; the accumulated crystals, and crystals which nucleated and grew from the trapped liquid in the pore spaces between them. Proportion of these two components in the final solidified rock is influenced by numerous processes which affected its crystallization, and provides basis for the division of cumulates into several types.

Cumulates are predominantly mafic to ultramafic, given by the fact that conditions in magma bodies of such compositions are more favourable for their formation (low viscosities and a perceptible density difference between liquid and solids supports efficient separation of crystals, Holness 2020). Nevertheless, cumulate rocks are also, although significantly less frequently, found in intrusions of more evolved compositions, such as syenites or granites (Upton 1996, Marks and Markl 2015).

II.1. Terminology

The term 'accumulative rock' has gradually begun to appear in context of igneous petrology after Bowen's work in 1928, in which he proposed crystal accumulation without remelting as a likely origin of certain ultrabasic rocks. In the following years, several observations supporting this idea have been made during detailed mapping of layered mafic intrusions (especially of Skaergaard intrusion in eastern Greenland and Rhum intrusion in Scotland) (Wager and Deer 1939), as some of the petrological relations and textural features found in these igneous bodies indicated the origin by crystal accumulation.

Expanding research on layered mafic intrusions led to a wider acceptance of crystal accumulation as the formation mechanism of certain gabbroic rocks, as well as some of the monomineralic rock types such as dunites or chromites. In response to that, the group of accumulative rocks has soon taken on a new name ‘cumulates’ suggested by Wager et al (1960), along with a set of textural criteria by which the rocks can be defined and interpreted in terms of their cooling history. Wager’s nomenclature of cumulate textures received a broad recognition and is, with certain modifications, still used today.

The entire nomenclature stands on the idea that after the initial accumulation of crystals (termed ‘cumulus crystals’, ‘primocrysts ‘or ‘primary precipitates’ in older literature), there are numerous ways in which the crystallization can be completed, and the variations in extend of each process taking place will produce important differences in the solidified rock. Wager (1963) distinguishes three stages in the formation of cumulate rocks, each having its impact on the final texture of the cumulate:

1. Homogeneous nucleation and growth of primocrysts, followed by their deposition.
2. Growth of cumulus crystals deposited on top of the pile via diffusive exchange with the bulk magma (a process referred to as 'adcumulus growth').
3. Crystallization of the interstitial (or 'intercumulus') liquid trapped in between the cumulus crystals, as the deposited crystals are buried deeper by an overlying pile of accumulated crystals.

The decisive factor here is the ratio between the relative extent of processes in the second and third stage of formation. If the cumulus crystals do not grow any further after their deposition (=there is no adcumulus growth) and all the trapped liquid crystallizes in pore spaces between them, the resulting cumulate rock will be referred to as 'orthocumulate'. If, on the other hand, the cumulus crystals proceed to grow after their deposition while nearly all the intercumulus liquid is excluded, the resulting rock will be an 'adcumulate'. The connecting link between these two extreme scenarios are 'mesocumulates', where some adcumulus growth took place, followed by the crystallization of trapped liquid (Fig. 1).

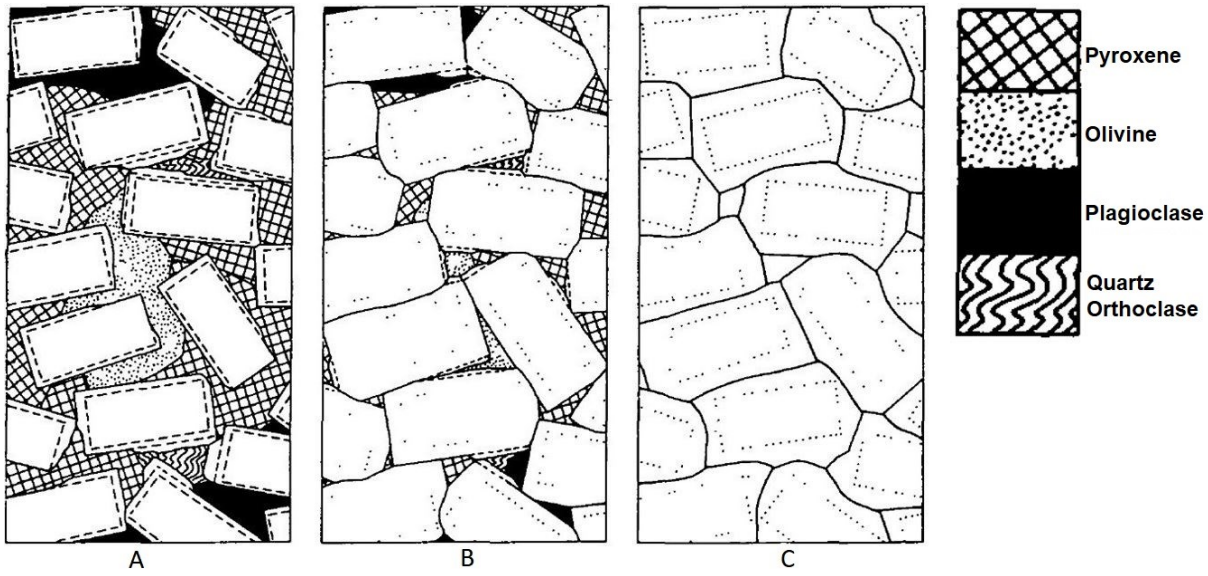


***Figure 1** - Diagrammatical sketch representing the three main types of cumulates, with plagioclase (white rectangles) as cumulus phase: A) orthocumulate; B) mesocumulate; C) adcumulate. From Wager et al (1960).*

II.1.1. Naming a cumulate rock

Wager et al (1960) suggested that cumulate rock should be named according to the cumulate crystals it contains, lined up from the most to the least abundant mineral phase. This method proved most useful during mapping of layered intrusions, because while the rock might have undergone some postcumulus changes which would affect its name in the traditional system, Wager's classification would consider the rock unchanged, which helps to precisely correlate layers with the same assemblage of cumulate minerals throughout the stratigraphy (Jackson 1967).

Its disadvantage lies in the fact that the name can become impractically long when the cumulate rock consists of more than two cumulate minerals. For mapping or drill core description, this can be avoided by using abbreviations for each cumulate. For example, when describing the stratigraphy of the Stillwater complex, PoC stands for a plagioclase-olivine cumulate (conventionally named troctolite) (Irvine 1982, McCallum 1996). However, as good as the cumulate names are for research purposes, the conventional rock names still dominate in published work and textbooks. It is, after all, more practical and pleasant for a reader to use familiar names such as norite, instead of 'plagioclase-orthopyroxene cumulate' (Irvine 1982).

II.2. Architecture of cumulates

The proportion of cumulus and intercumulus material in a cumulate rock is its most pivotal characteristics, as it provides an important insight into the conditions under which it formed and the processes which shaped its texture. However, it is important to keep in mind that determining the proportions of cumulus crystals and intercumulus material is not always straightforward. Cumulate rocks exhibit enormous diversity in their characteristics, and so the traits which can potentially help us establish the amount of intercumulus material can sometimes be unclear or completely obscured (McBirney and Hunter 1995).

The cumulate crystals are typically euhedral to subhedral, with either no compositional zoning, or compositionally constant cores with overgrowth rims which crystallized from the trapped liquid (Wager et al 1960). Their composition corresponds to the bulk magma composition at a time of their growth, meaning that the mineral assemblage of primocrysts and the composition of their unzoned cores changes over time as the magma in the intrusion evolves (Holness 2020). A good example of this can be found in Skaergaard layered intrusion, where the composition of the cores of plagioclase primocrysts becomes gradually more albite-rich with increasing stratigraphic height (Humphreys 2009).

The overgrowth rims on cumulus crystals are commonly zoned (Wager 1963). The zoning is usually normal, but can be reverse or oscillatory, again depending on the processes taking place during their growth. If the compositional zoning is not developed or is obscured by later processes, the overgrowth rims can be almost indiscernible from the crystal they surround (Campbell 1987).

The intercumulus crystals are anhedral and often poikilitic in habit, and typically compositionally zoned (Barnes et al 2016). The degree of their zonation depends on the mobility of interstitial liquid; if the liquid is trapped and immobile, it will undergo a progressive fractionation and resulting intercumulus minerals will be distinctly zoned. If however the permeability of the mush is high enough for a compositional convection to take place within it, the resulting mineral zonation can be limited or completely obscured (Sparks et al 1985, Campbell 1987).

II.2.1 Orthocumulates

Orthocumulates are a cumulate type containing more than 20% of crystallized intercumulus liquid, with very little or no adcumulus growth affecting the primocrysts. The upper limit of the amount of interstitial material in an orthocumulate is not strictly defined, as it is controlled by the packing of cumulate crystals which can be of various sizes and shapes (Holness 2020). Maximum initial porosity of a crystal mush has been modelled by numerous analogue experiments, for example using water and aluminium tablets as imitation of feldspar crystals, and the results show that the initial porosity could commonly be within the range of 40 to 60 vol.% (Wager et al 1960).

Pure orthocumulates are considered an extreme scenario and are found very rarely in nature (Irvine 1982), as the primocrysts typically undergo some adcumulus growth before the system becomes isolated from the bulk magma. For this reason, the definition of orthocumulates has been mitigated to include rocks which slightly deflect from the ideal case.

Illustrative examples of this cumulate type can be found in the Lower zone A of the Skaergaard intrusion, as these rocks have been used as exemplary case by Wager et al (1960) in their work which laid the groundwork of cumulate terminology (Fig. 2). These rocks are composed of olivine and plagioclase primocrysts, surrounded by large poikilitic pyroxenes which crystallised from the interstitial liquid together with small amounts of Fe-phases and micropegmatite (which represents the last remaining, highly fractionated droplets of liquid) (Wager 1963).

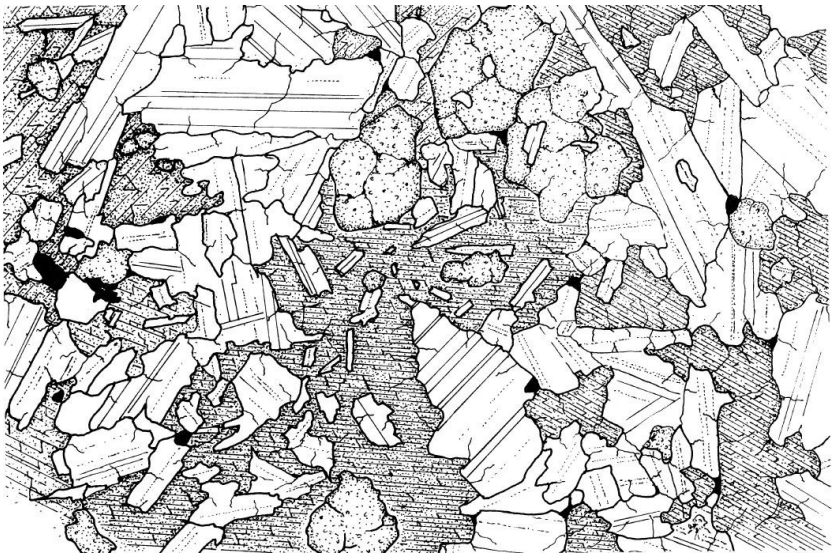


Figure 2 *– Outlined thin section of an orthocumulate from Skaergaard's Lower zone A. The rock consists of plagioclase (unpatterned) and olivine (stippled) primocrysts, surrounded by large pyroxene oikocrysts (striped) and a small amount of Fe-oxides (opaque). From McBirney and Hunter (1995).*

II.2.2. Adcumulates

Adcumulates are the opposite extreme case to orthocumulates, meaning that the cumulate minerals were affected by extensive adcumulus growth during the solidification of the cumulate. The rock is made up almost entirely of large primocrysts, which are nearly uniform in size and either show very little compositional zoning or are completely unzoned. If the primocrysts are made up of a single mineral phase (for example only plagioclase), the resulting cumulate rock will be almost purely monomineralic (Fig. 3). Intercumulus material is scarce (less than 5 vol.%), as most of the intercumulus liquid has been forced out of the mush during the formation, leaving only small residues preserved as anhedral and typically zoned crystals (Wager et al 1960, Wager 1963). Adcumulates are in most cases depleted in incompatible elements, as a result of expulsion of interstitial liquid (Latypov and Chistyakova 2014).

Even though 'perfect' adcumulates with no intercumulus material are a very rare occurrence, the rocks which meet the 5% criteria are not uncommon and can be found in majority of layered mafic intrusions (Irvine 1982, Morse 1986). As a representative example, anorthositic gabbros from the Layered series of Skaergaard intrusion are often used to show a typical adcumulate texture. The rock consists of nearly uniformly large, unzoned plagioclase primocrysts, accompanied by a small amount of intercumulus pyroxene and olivine crystals scattered around in between the larger grains. The Fe-oxides commonly found as the intercumulus material in cumulates are missing, as well as any of the more evolved phases (Wager 1963).

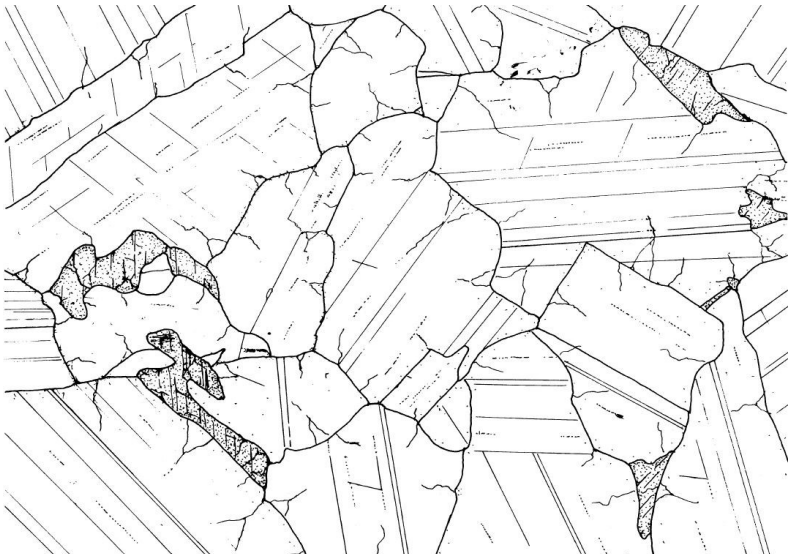


Figure 3 *– Outlined thin section of an anorthositic gabbro from Skaergaard's Layered series, representing a typical adcumulate texture – The rock is made up of uniformly large plagioclase crystals, with very little interstitial material. From McBirney and Hunter (1995).*

II.2.3. Mesocumulates

Mesocumulates are a transitional type between the two extreme cases – orthocumulates and adcumulates, meaning the primocrysts underwent some adcumulus growth, but not enough to expel the vast majority of intercumulus liquid (Wager et al 1960). Although this type should be the most common one in nature, the term 'mesocumulate' is not frequently used in literature. This is perhaps due to the fact that the nomenclature is often used loosely so that the end members include a wider range of cases, and also because the extreme types naturally receive far more attention.

II.2.4. Heteradcumulates

Heteradcumulates are made up of unzoned primocrysts surrounded by a significant volume of large, mostly unzoned poikilitic crystals. As suggested by the name, this cumulate type is more related to adcumulates to which they can be linked by a similar formation mechanism (Wager et al 1960). In this case, the crystals nucleating from the pore liquid grow alongside with cumulate crystals by a diffusive exchange with the bulk magma, resulting in unzoned poikilitic crystals (termed as 'heterad material') and primocrysts. The heterad material thus might have the same compositions as primocrysts in an adcumulate formed from the same magma.

Heteradcumulates are notoriously difficult to distinguish, as the zonation of intercumulus material or rims around the primocrysts in orthocumulates can sometimes be underdeveloped or obscured by later processes, causing the two types to look roughly similar. As a result, there are a few ongoing debates regarding some of the cumulate rocks found in mafic intrusion which may or may not have been incorrectly described (McBirney and Hunter 1995).

II.2.5. Crescumulates

Also known as harrasitic cumulates, this type is presumed to form predominantly in-situ on vertical or steeply dipping walls of the intrusion. It is characterised by long dentritic crystals which grew parallel to a thermal gradient (and perpendicular to igneous layering) in a significantly undercooled system. This trait distinguishes them from typical in-situ formed cumulates, which commonly do not have any shape preferred orientation (SPO) of their primocrysts (Wadsworth 1961, Holness 2020).

Such cumulates can be found for example in the layered series of the Rhum intrusion (Fig. 4) (Wager et al 1960). The large, elongated olivine primocrysts, branching upwards perpendicular to the base of the layer, are surrounded by poikilitic, mostly unzoned plagioclase and clinopyroxene intercumulus crystals outstretched in the same direction. A remarkable example can be also found the in Skaergaard intrusion (in one of the xenolithic blocks, interpreted as a part of the former roof which tore off and sank down) where the crescumulates exhibit rhythmic layering created by alternating between the layers with plagioclase and pyroxene primocrysts (Wager and Brown 1968).

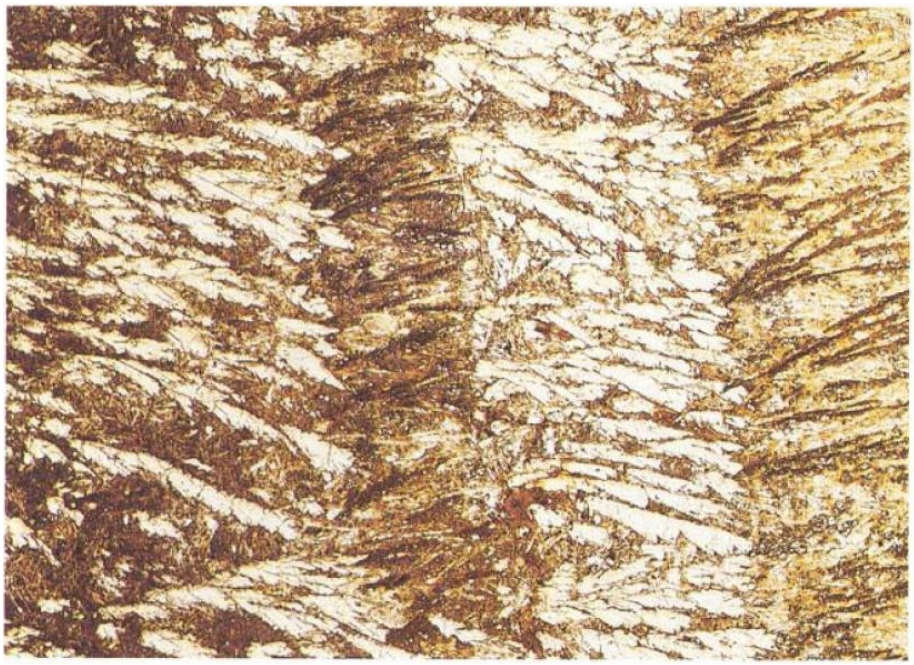


Figure 4 *- An example of crescumulate texture in dolerite dike, from North-western Skye (Scotland). Starting from the left side, the first and third layer are dominated by elongated, branching augite crystals, while the second and fourth contain mainly branching olivine. 8x magnification, parallel polars. From Mackenzie et al (1982).*

II.3. Inaccuracies in cumulate definition and terminology

The nomenclature and interpretation of cumulate textures by Wager et al (1960) proved exceptionally useful, as it provided petrologist with a simple and logically structured terminology, a tool for research and clarification of some igneous structures. However, as the knowledge on the subject began to progress further, some researchers started to question certain aspects of Wager's work.

The first small hitch appeared when the research has shown that in case of ultramafic bodies, plagioclase can often be less dense that its parental liquid and therefore it should float, instead of settling down to the floor of the intrusion (Bottinga and Weill 1970). Given that Wager et al (1960) used plagioclase cumulates from Skaergaard's Lower zone as illustrative examples, the discovery sparked questions regarding the formation of these rocks and whether crystal settling was a necessary process to begin with. In response, some scientists even chose to avoid the usage of Wager's nomenclature altogether, as to prevent any misleading conclusions (McBirney and Noyes 1979).

A serious challenge to the theory of cumulate formation was presented by Campbell (1978), who argued that in-situ crystallization and growth could not only be another possible formation mechanism, but also equally important as crystal settling. Many field examples of cumulates formed where the crystals could not have settled (or floated), for example on steep walls of layered intrusions (McBirney and Noyes 1979), supported this view which later gained recognition. As a result, the definition of cumulate rocks needed to be modified, and so Irvine (1982) proposed to redefine the

word ‘cumulate’ so that crystal settling is ‘possible but not essential process’ in the origin of the rock to which we apply it.

Another problematic (and still unresolved) topic is dividing the cumulates into groups such as ‘orthocumulates’ based on Wager’s textural criteria, as they do not consider any post-cumulus equilibration and alterations (Hunter 1987). Although the usage of these terms is very practical for mapping and research purposes, it is important to keep in mind that the textures we observe today could have been extensively altered, and so the features based on which the cumulate types are commonly distinguished can be completely obscured. For example, if a melt percolated through the mush during the solidification, then the traditional methods for assessing the amount of trapped liquid in a cumulate rock will produce misleading results, as majority of these methods do not consider open system behaviour (McBirney and Hunter 1995). The inevitable question then is whether it is appropriate to use Wager’s nomenclature if we are unable to reliably determine the fundamental properties it was based on.

As McBirney (1995) pointed out, although the formation mechanisms used to formulate the cumulate theory and nomenclature are undoubtedly valid, so too are other processes which were not considered at that time. At present, most petrologist recognize the problems with cumulate terminology, yet no better alternative has been presented and the terms continue to be routinely used, while keeping the uncertainties in mind.

III. Geological setting: Layered intrusions

Mafic cumulates can be found across many different types of intrusive bodies, ranging from small sills less than ten centimetres thick, to immense magma chambers reaching up to hundreds of kilometres in diameter (Chistyakova and Latypov 2014, Cawthron 2015). The occurrence of cumulate rocks thus depends on cooling history of the body and processes active during its lifetime, rather than directly on its size.

Layered mafic intrusions represent large magmatic bodies of mainly basaltic composition, typically of lopolithic to narrow funnel-like shape and variable dimensions (Fig. 5), reaching up to tens of thousands square kilometres of exposed area (Cawthron 2015). These intrusions have been a subject of extensive studies especially in the last six decades, partly for their unusual mineralization and ore-mining potential, but mainly for their stunningly detailed records of crustal magma chamber processes, such as crystal accumulation and growth, fractional crystallization, convection, or chamber replenishment, which are preserved in their structure (Scoates and Wall 2015).

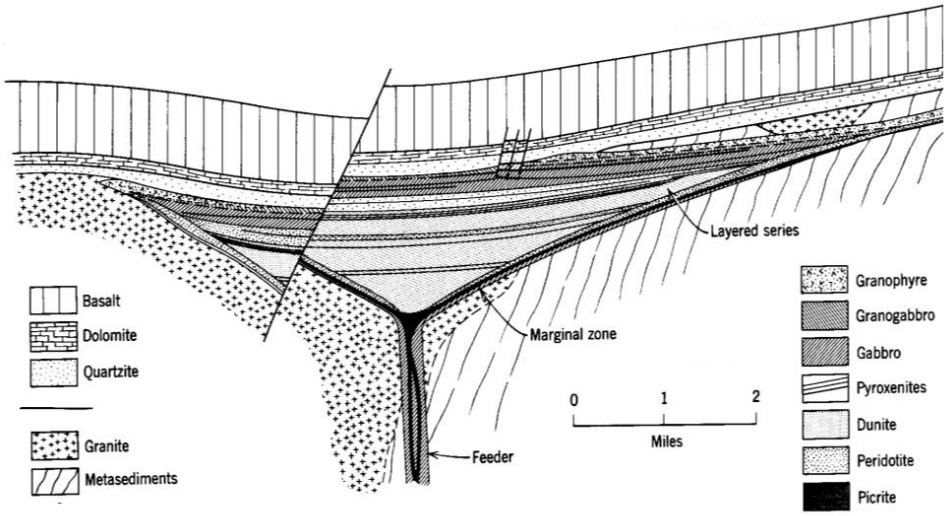


Figure 5 – Nearly ideal lopolithic shape of layered mafic intrusion represented by Muskox intrusion in Nunavut, Canada. From Irvine and Smith (1967).

Although broadly studied, the topic of layered mafic intrusions is still an area of major debate, and only very few agreements have been reached concerning the origins of their structures, such as their signature layering, and the character of processes they preserve (Holness 2017, Nielsen and Tegner 2017). Nevertheless, as a fountainhead of a vast variety of cumulate rocks, layered intrusions provide an opportunity to resolve the fundamental questions regarding the nature and formation of these rocks.

III.1. Structural setting and formation

There is no single specific geological setting in which layered intrusions occur, and so their geological background and conditions under which they form cannot be strictly defined. However, it has been observed that most of the known intrusions are found near both successful and failed rift systems, and so the parts of crust undergoing extension seem to be favourable for their emplacement. In regard with that, layered intrusions are often being put into context with large igneous provinces as a possible magma storage, feeding the extensive lava flows erupted in the early stages of rifting (Ernst et al 2017, Bryan and Ernst 2018).

The oldest layered intrusions have been dated to late Archean, approximately 2.75 Ga. (Campbell 1982) The fact that no older intrusions have been discovered might suggest that physical properties of crust prior to this time had not yet been ideal for the emplacement of such large magma chambers. On the opposite side of a timeline, youngest exposed intrusions are Tertiary.

Parental magmas of layered intrusions are variable in composition, including komatiites, high-alumina basalts, alkali basalts and mainly tholeiites. The MgO content of these magmas is commonly placed within the range of 10-14% (Campbell 1982). However, majority of intrusions have been replenished by several separate magma pulses during their lifetime with progressively changing compositions due to differentiation of their source, making the description of parental magmas difficult and often inaccurate (Jenkins and Mungall 2018). Moreover, there are quite a few different approaches to decoding the composition, such as analysis of chilled margins or dykes crosscutting the country rocks in close proximity to the intrusion, and there does not seem to be any agreement on which approach produces the most accurate results (Keiding et al 2010, Jenkins and Mungall 2018).

III.2. Layers and layering

The name “layered intrusion” directly refers to a signature layering of cumulate rocks found in these igneous bodies. Is it defined as an overall structure developed by a combination of separate layers, forming a sheet like cumulates, each being compositionally or texturally distinctive from its neighbours. (Namur et al 2015) Individual layers laid on top of each other in a stripe pattern make up a sequence of rocks in a vast variety of combinations, creating the layered structure (Fig. 6).

Within a single intrusion we can observe variations in every possible characteristic that a layered sequence, or even a single individual layer can have. Layering itself is often planar, but it can as easily form many other structures such as waves, convoluted pattern with layers blending into one another, troughs etc. Moreover, layered structure is not always prominent and can be very poorly defined, to the point where it is almost impossible to distinguish it with a naked eye.

Individual layers occur as uniform, as well as stratigraphically variable (Wager 1963). They often show diverse types of grading, such as textural (based on crystal morphology), grain size grading, modal (melanocratic/leucocratic) or cryptic (chemical) (Wager and Deer 1939, Namur et al 2015). Layers are also described in terms of the nature of their contacts, which can be gradational or sharp, their thickness, ranging from microscopic to hundreds of meters, and their form (e.g., planar, convoluted, antiformal etc.) (Irvine 1982). All these characteristics can be easily observed and described in majority of layered intrusions; however, their origin and mechanisms of formation remain unclear in many cases.

Figure 6 *- Field photograph of a layered structure visible on a steep cliffside near the upper border of Skaergaard's Middle zone. The three most prominent layers, made up of a plagioclase-rich cumulate, are known as the 'Triple group' and represent an important stratigraphic marker. Each layer is approximately 5 to 7 meters thick. Modified after Holness (2017).*

Phase layering

A major type of layering based on which are layered intrusions usually mapped and subdivided into zones and series. It is defined by the first appearance or disappearance of a specific cumulate, or a particular cumulus mineral throughout the stratigraphic sequence of the intrusion (Irvine 1987).

Rhythmic layering

Rhythmic layering is characterized by a systematic repetition of distinctive layers of the same kind, or their specific sequence, and can be subdivided into macro and microrhythmic type depending on the thickness of individual layers. Macrorhythmic layering is characterized by greater layer thickness, reaching up to first units of meters. A well described and instructive example of macrorhythmic

layering can be found in the Lower and Middle zone of Skaergaard intrusion where the layered sequence is made up of melanocratic and leucocratic layers, each 3 to 8 meters thick (Irvine 1982).

In microrhythmic layering, bands are typically fine scaled, most commonly ranging from microscopic to centimetres in thickness. Visually, the structure of microrhythmically layered cumulate rocks closely resembles tidal layering, a sedimentary structure created by alternate flow of water on tidal flats (Nicols 2009). A great example of rhythmically layered cumulates can be found in the Lower banded series of Stillwater complex. The rocks of the Norite I. zone exhibit prominent fine-scale layering (Fig. 7), created by alternation of pyroxene-enriched and pyroxene-depleted bands, approximately 3 to 10 cm thick (Boudreau 2011).

Figure 7 *- A field example of microrhythmic layering from Stillwater complex. Layering in norites is defined by depletion and abundance of pyroxene on a scale of units of centimetres. Hammer for scale. From Driscoll et al (2015).*

Cyclic units

One cyclic unit is made up of several cumulate layers with gradually changing composition throughout the unit, following the fractional crystallization order of magma (Jackson 1970). Therefore, a complete cyclic unit starts with olivine rich cumulates at its base, then gradually (or abruptly) passes into pyroxenite, followed by anorthosites at the top. This cumulate sequence can repeat itself many times, often with some variations in geochemistry and thickness between successive cycles. Cyclic layering is

often interpreted as a record of magma chamber replenishment, with each cycle representing a new pulse of magma entering the chamber (Irvine 1987).

III.2.1. Layer forming processes

As previously mentioned, the majority of questions regarding the origin of igneous layering has not yet been fully resolved. Given the incredible variety in layering types and forms, it is no surprise that many different explanations have been offered for their formation. Most of the suggested processes can be separated into two main categories: Dynamic and non-dynamic. However, it is important to note that some of these processes may have both dynamic and non-dynamic components to them, and so this division may not always be appropriate (Irvine et al 1998).

Dynamic processes incorporate the internal movements of melt and crystals within the magma chamber. Crystal settling (Wager and Brown 1968), magma currents (Irvine et al. 1998, McBirney and Nicolas 1997), convection (McBirney and Noyes 1979), and mixing (Holness et al 2007, Cawthron and Ashwal 2009), or flow segregation (McBirney and Nicolas 1997) are good examples of such processes and are commonly associated with the formation of layering. Some of these mechanisms, such as density currents, are especially effective near the margins of the chamber (Namur et al. 2015).

Non-dynamic processes take place during the crystallization and do not involve any large-scale movement of magma or crystals within the magma chamber. Most of these processes can be attributed to condition changes during crystallization (e.g., changes in pressure and oxygen fugacity, or nucleation and growth rate fluctuations; Wager and Brown 1968, McBirney and Noyes 1979, Naslund and McBirney 1996) or equilibration and self-organization in the crystal mush (e.g., dissolution-precipitation, mush compaction, etc.; Boudreau and McBirney 1997, Boudreau 2011).

III.3. Petrography

With layered intrusions showing incredible diversity in all aspects, it is crucial to mention that to portray their rock types and common textures in such a way that would be fitting to every layered intrusion is an impossible task. And so, all characteristics further described should be accounted only as common, not universal.

First and foremost, it is important to note that the mafic rocks making up the intrusions are often accompanied by intermediate and felsic types, such as granites, diorites and syenites, as is the case for Bushweld and Kiglapait intrusions (Morse 2015, Cawthorn 2015). These rocks are mostly situated at the stratigraphic top of the intrusion and are subordinate by volume to mafic rock types present in the complex. Both granitic and syenitic rocks are predominantly medium to coarse-grained (although fine-grained varieties are not uncommon) and are generally more texturally and compositionally homogeneous compared to their mafic neighbours (Cawthorn 2015). Layering is observed in granitic

rocks rather rarely, while it is not uncommon in syenitic parts of layered intrusions (Marks and Markl 2015). One of the best examples of layered syenites can be found in the Ilímaussaq igneous complex in Greenland, an extraordinary layered intrusion composed predominantly of peralkaline syenitic rocks (Upton et al 1996, Marks and Markl 2015).

III.3.1. Mafic cumulates

Mafic and ultramafic cumulates represent a vast majority of rocks commonly found in layered intrusions. Characteristics of different types of cumulate rocks have already been discussed in detail in the previous chapter, and so for the purpose of describing layered intrusions from petrological point of view, it is merely important to mention again that cumulate rocks are further divided into several types; orthocumulates, mesocumulates, adcumulates, etc., based on the processes involved during their formation and the resulting propotion of cumulus and intercumulus material in the solidified rock. Distribution of these cumulate types in layered intrusions varies from one intrusion to another and does not seem to follow any strict and generally applicable pattern, however, several observations regarding their distribution have been made (Campbell 1982):

- The textures of layered intrusions can be both progressively more adcumulate towards the roof of the intrusion (Skaergaard, Kiglapait, etc.) or on the contrary, the proportion of intercumulus material can gradually decrease from the roof downward, creating a reverse case (Jimberlana intrusion).
- Individual layers often become more orthocumulate in texture toward the margin of the intrusion. This phenomenon is usually attributed to the direction of heat loss.
- With an arrival of a new cumulus phase, the bottom part of the newly created layer will be in most cases more adcumulate that the top of the older layer underneath it (Morse 1979).

As mentioned before, layered intrusions are subdivided based on phase layering aka the first appearance/disappearance of a cumulus phase. Phase layering thus creates a stratigraphic sequence unique to every intrusion and no main similarities in its arrangements have been observed. What however does repeat itself and can be attributed to every intrusion, are the main rock types making up the zones and subzones.

In general, the most common are gabbronorites, norites and olivine gabbros, with a variable grain size throughout the intrusion ranging from fine-grained to pegmatitic. These cumulates make up the largest volume percentage of many stratigraphic zones, such as the Main zone of the Rustenburg Layered suite in Bushveld (fine-grained norite and gabbronorite) or the Upper banded series of Stillwater complex (olivine gabbro and gabbronorite) (Cawthron 2012, Jenkins and Mungall 2018). Moreover, their fine-grained varieties are commonly found in the marginal zones, aka the chilled margins of the intrusions.

Another important cumulate type occurring in the vast majority of intrusions are troctolites. These greyish rocks are composed merely of plagioclase and olivine grains, both of which are often easily distinguishable thanks to their contrasting colours (light plagioclase surrounding the dark olivine grains). Troctolites are most commonly medium to coarse-grained. In some cases, such as the Kiglapait intrusion in Canada, troctolite can be a predominant rock type making up more than half of the complex (Morse 2015), while in many other intrusions it can be a rare occurrence.

Monomineralic cumulates represent a whole group of rocks commonly found throughout these igneous complexes. These rocks are composed mainly of a single kind of cumulus mineral (more than 90%), with very little or none interstitial space left to be filled with other mineral phases. The most frequent monomineralic rock types occurring in layered intrusions are dunites, anorthosites, bronzitites, and chromitites, each very variable in their properties such as grain size, percentage of minor mineral phases, etc. As an example, Critical zone of eastern Bushveld complex offers one of the most instructive cases of a layered sequence created purely by monomineralic cumulates, with thin layers of massive chromitite placed within anorthosite cumulates (Fig. 8), creating a striped pattern which can be followed for many kilometres around the perimeter of the intrusion (Latypov et al 2017).

Figure 8 *- The alternation of anorthosite and chromitite layers in the Critical zone of Bushveld complex. From Boudreau (2011).*

III.3.2. Mineral deposits

The layered intrusions are well known amongst geologist for their unique mineralization. They contain world class deposits of many essential metals such as platinum group elements or PGE (Ru, Rh, Pd, Os, Ir, Pt), vanadium and chromium, for which they represent the only economically profitable resource (Lee 1996).

Metalliferous mineralization of layered intrusion includes ilmenite, magnetite, chromitite, gold and several kinds of metal sulphides (Lee 1996). Their deposits are distributed unevenly throughout the intrusion, occupying only certain narrow layers predominantly in lower to central parts of the body. PGE do not occur as independent ores and so are mostly found in association with mineral phases previously mentioned, commonly in combination with Ni and Cu, or some of the metalloid ligands such as Te, Sb, and As (Maier 2005, Sahu et al 2020). Average grade of ores mined primarily for PGE ranges from 5 to 15 ppm (Sahu et al 2020), while in case of other elements such as chromium, the grade is typically much higher.

In case of PGE mineralization, the layers in question are referred to as reefs. These stratiforms can often be traced throughout the whole intrusion and represent important stratigraphic markers. PGE deposits can be further divided into several types based on their stratigraphic position and rock types they are commonly associated with (Maier 2005):

- "Contact" PGE reefs – PGE mineralization occurring nearby the contact with the country rock, at the sidewall or at the base of the intrusion. The PGE are commonly hosted by gabbronorites or pyroxenites. These reefs typically have higher sulphide abundances compared to other types (between 5 to 10 percent in average), but the contents of PGE tend to be lower. The base metal sulphides are mostly pyrrhotine, pentlandite and chalcopyrite.
- PGE reefs in lower to middle parts of the intrusion – These deposits are typically characterized by wide zones (tens to hundreds of meters) of disseminated sulphides, hosted mostly by pyroxenites and peridotites. For example, the well-known Merensky reef of the Bushweld complex falls into this category.
- PGE associated with massive chromitite layers – These reefs are generally sulphur-depleted in comparison to other types, which is often a cause of a heated debate regarding the origin of their mineralization. PGE concentrations vary strongly from layer to layer in a single intrusion. The Bushweld complex can again be used as an illustrative example; The UG1 and UG2 chromitites of the Upper group contain 1 ppm and 7 ppm respectively (Latypov et al. 2017) which makes only one of the layers economically lucrative.
- PGE reefs in the upper parts of the intrusion – unlike other reef types, these deposits are often hosted in gabbro-diorites and Fe-Ti oxide rich layers. They are not nearly as common as

previously mentioned types and are mostly economically nonprofitable. Nevertheless, these layers do contain considerable PGE concentrations compared to surrounding rocks.

Table 1** - **Comparison of mineralization in a variety of layered intrusions. After Lee (1996).

Layered complex	Pt>Pd	Pt≈Pd	Pt<Pd	Au	Cu	Ni	Co	Cr	Ti	V
Bushveld	◊	◊	◊	◊	◊	◊	◊	◊	◊	◊
Stillwater			◊		◊	◊		◊		
Great Dyke		◊			◊	◊		◊		
Windimurra			◊							◊
Munni Munni			◊		◊	◊				
Muskox		◊			◊	◊		◊		
Skaergaard			◊	◊						

The ◊ symbol marks an occurrence of commodity which is, has been, or has the potential to be actively mined, or is obtained as a by-product. Notable is the range of PGE mineralization in terms of Pt/Pd ratio – majority of PGE occurrences are Pd-dominant, with the exception of Bushveld complex.

IV. Formation of cumulates I: Accumulation

Formation of a cumulate rock can essentially be viewed as a three-part process. During the first stage, primocrysts form a crystal mush either by their in-situ nucleation and growth, or by redistribution from their original growth site followed by accumulation in a different part of the intrusion. Throughout the second stage, primocrysts continue to grow from the interstitial liquid which at this point still maintains communication with the bulk magma, and therefore is able to retain a similar composition. As the growth progresses, it causes the permeability of the mush to gradually decrease (which can be further assisted by an expulsion of the interstitial liquid due to compaction), until it consequently traps the remaining interstitial melt, and the formation process enters a third stage: Crystallization of intercumulus material (Wager 1963, McBirney and Hunter 1995, Holness et al 2017).

The focus of this chapter will primarily be on the first stage of formation and the processes driving both redistribution and in-situ nucleation and growth of crystals.

From the very beginning, the formation of cumulates was attributed to gravitational settling of crystals, as the structures observed in some mafic igneous bodies (namely layering or density-sorted grains) showed conspicuous similarities to common features found in sedimentary rocks (Wager and Deer 1939, Wager and Brown 1960). In fact, this analogy which laid the foundations of Wager's cumulate theory has been so compelling that it has not been seriously opposed for more than ten years. However, as the research of layered mafic intrusions deepened, more structures which could not have been easily explained by crystal redistribution were found (for example crescumulates formed on steep margins, or microrhythmic layering). Consequently, the idea of in-situ nucleation and growth gained relevance and took the cumulate theory by storm.

Today, it is generally acknowledged that the role of neither of these processes should be neglected as layered intrusions mostly show evidence for the operation of both redistribution and in-situ nucleation, although the latter is often considered to be the dominant process (Holness 2020). Opinions on the extent to which crystal redistribution contributes to the assemblage of crystal mush differ greatly between researchers, especially when discussing crystal settling; while some still attribute key importance to it (e.g., Eales 2000), others choose to avoid it as much as possible (e.g., Latypov et al 2017). As a result, one cumulate rock or one layered sequence commonly has multiple different interpretations and the literature is full of polite, but firm disagreements over some of the most elementary aspects of their formation.

IV.1. Cumulates formed by crystal redistribution

The idea of cumulate formation by crystal redistribution suggests that the primocrysts nucleated and grew in one part of the intrusion and later were transported to the place of their deposition and final solidification. Transport of grains throughout the chamber can be achieved by a number of processes, such as convection, density currents, flotation, etc. Therefore, although majority of cumulates formed by crystal redistribution is found on the floor of the intrusion, redistribution does not exclusively imply a downward movement.

There are only a few common features which can be used to generally describe these cumulates. Each process possibly involved in their formation can influence the final structure of a rock in different ways. Therefore, it is once again important to stress that the characteristics further described should be taken only as the most common or ideal, not as a generally applicable rule. Also, some of the field examples used to illustrate them could have been given various other interpretations by different researchers, not always in association with crystal redistribution (such as a dispute between Holness et al (2017b) and McBirney and Noyes (1979) regarding Skaergaard trough layering).

Typical placement of cumulates formed by crystal redistribution is either the floor, or occasionally the roof of the intrusion (Irvine 1982). Transport of grains towards the roof requires special conditions and therefore the so-called 'flotation cumulates' do not form as frequently (Namur et al 2011). Deposition on the floor mostly results in the formation of horizontal or gently dipping cumulate layers, which can display characteristics similar to sedimentary features such as density sorting, normal or reversed gradation, or imbrication. Shape preferred orientation of primocrysts, most commonly of elongated plagioclase grains, is another frequent attribute. Unlike primocrysts, interstitial material does not show any preferred orientation as it developed after the accumulation (Holness 2020).

IV.1.1. Crystal settling

Crystal settling can be considered as an analogous process to sedimentation. Following their nucleation in one part of the chamber, crystals will begin to settle down towards the floor of the intrusion similarly to grains sinking in an aquatic environment. The nature of both environments is, of course, drastically different, and yet the structures observed in rocks resulting from these processes closely resemble each other.

The settling velocity (V) of a crystal sinking down to the floor of the intrusion can be defined by Stokes' law:

$$V = \frac{2}{9}\eta g \Delta \rho r^2, \qquad (1)$$

where r stands for the radius of the grain (squared), η for the viscosity of the surrounding magma and $\Delta\rho$ for the density difference between the solid and the liquid. As a resulting effect, mineral grains with various sizes, densities and shapes will have different settling velocities, which in turn causes density-sorting, and modal and grains-size gradation in accumulated layers (Namur et al 2015).

In an ideal layered sequence formed by crystal settling, the high-density mineral phases such as olivine or pyroxene should be concentrated at the bottom of the sequence, while lighter minerals like plagioclase form the uppermost layers (Naslund and McBirney 1996). Meanwhile, the minerals grains of the same type should be sorted throughout the layer in terms of size, with the coarsiest grains at its base (Campbell 1978, Higgins 2002). Examples of layering exhibiting such features are not particularly rare and can be found in a number of layered intrusions. In case of the ultramafic layered complex on Duke Island in Alaska, Irvine (1974) describes a rhythmically layered series consisting of olivine and pyroxene cumulates in which the finer olivine crystals are found at the top, and both minerals are size graded within each layer. As another example, the cyclic units of Muskox intrusion in Canada are also often interpreted as a sequence formed by a crystal settling process. Each cyclic unit consists of a basal dunite layer, followed by harzburgite and then othropyroxenite as the uppermost segment, with each layer exhibiting grain size gradation (Irvine and Smith 1967).

The view on crystal settling had significantly changed over time, as some aspects of the original cumulate theory regarding the process are presently considered problematic. For instance, the theory presumes homogeneous nucleation near the roof of the intrusion from where the crystals begin to settle as individual grains. Later however the opinions on where and how the crystal nucleate began to strongly vary, and more and more researchers spoke against a widespread homogeneous nucleation in large, slow cooling intrusions (Campbell 1978).

Another important aspect to consider regarding the effectivity of crystal settling is convection. In the past, studies indicated that high convective velocities in mafic and ultramafic chambers could exceed the settling velocities by up to several orders of magnitude and thus convection should be able to keep the crystals in suspension, effectively inhibiting crystal settling unless considerable aggregation of crystals occurs (Sparks et al 1985). These conclusions then led to the assumption that efficient crystal settling in large magma chambers can take place only within separate, thin layers of liquid resulting from processes such as double-diffusive convection, chamber replenishment, or density currents. While it is true that crystal settling in such thin layers is indeed most likely effective, more recent studies and numerical models show that large scale convection can contribute to crystal settling just as well. While being carried by the flow of the convection cell, the grain in suspension can fall out of the current as it moves sideways above the chamber floor and accumulate there, forming a crystal mush (Verhoeven 2009).

Nevertheless, even though crystal settling is no longer considered as the main formation mechanism of cumulate rocks, plentiful field and experimental evidence of its operation shows that it is not negligible either.

IV.1.2. Crystal flotation

Crystal flotation is essentially an opposite process to crystal settling (Eq. 1), characterized by an upward movement of grains towards the roof of the intrusion resulting in the formation of cumulate layer accumulated from top to bottom. Such process requires very specific conditions in which the density of the surrounding liquid exceeds the density of the mineral grains. It is presumed that these particular circumstances do not frequently occur in mafic intrusions, and so the field examples of "flotation cumulates" are scarce.

According to numerous studies (e.g., Botinga and Weill 1970, Campbell et al 1978), an efficient crystal flotation can be achieved only with plagioclase grains in a primitive, ferrobasaltic melt. In addition, there may only be a small time window during which flotation is effective, as the magma along with the composition of plagioclase progressively evolves and the density contrast gradually decreases and inverts. With the magma evolving and the chamber possibly being replenished a number of times with fresh melt, there may be several episodes of plagioclase flotation and settling during the life of one intrusion (Namur et al 2011).

There are a few layered mafic intrusions, such as Skaergaard in Greenland, whose parental liquid compositions are assumed to have been suitable for an efficient crystal flotation (Botinga and Weill 1970), however, so far only a single intrusion has been found which exhibits a demonstrable example of a flotation cumulate. The Sept Îles ultramafic intrusion, located on the east coast of the Quebec province, is dominated by a basal series of troctolites and layered gabbros with an approximately 100 to 500 meters thick layer of anorthosite at the roof of the chamber. This anorthosite is interpreted as a flotation cumulate formed during the early stages of differentiation, before the melt reached the saturation of Fe-Ti oxides and the onset of their crystallization significantly decreased the density of the melt (Namur et al 2011). Anorthosite of identical compositions can also be found scattered around as centimetres to tens of meters-sized, slab-like blocks throughout the layered series, which represent former parts of the roof which torn off and sank down into the cumulate layers on the floor (Fig. 9).

IV.1.3. Density currents

Density currents in magma chambers are another phenomenon with an equivalent amongst sedimentary processes. In an aquatic environment such as continental shelves, higher concentrations of suspended sediment compared to surrounding water can easily become unstable on a sloping ground, perhaps due to the effects of earthquake or stormy weather, causing the suspension to slide down the slopes of the shelf onto the abyssal plains (Nichols 2009, Paull et al 2018). Similar process in magma chambers can be initiated by instability of a crystal mush forming on steep or vertical walls. Density currents are often a recurring process and thus are able to transport significant loads of crystals, making them an important contributor to formation of cumulates by crystal rearrangement (Irvine 1974).

Fragile crystal framework on the outer edge of cumulates forming on the walls of the intrusion can be disrupted by convective currents or collapse under its own weight. The possible trigger then sets the crystal-laden suspension into an avalanche-like motion, gathering even more material as it erodes the surface over which it flows, until it loses energy by descending onto a flat surface. Reaching the floor of the intrusion, the current spreads out to form subhorizontally layered cumulates (Irvine 1982).

Figure 9 – Autoliths from the Sept Îles intrusion in Québec, Canada. The anorthosite blocks formerly attached to the roof are now surrounded by rhythmically layered gabbros of the Megacyclic unit II in the Layered series. From Namur et al (2011).

Cumulates formed by density currents share a number of possible similarities with cumulates formed by crystal settling, such as modal and grain-size grading throughout the layered series resulting from one current deposition. In addition to that, layers often display other sedimentary traits characteristic for a deposition from a flowing medium, like cross-bedding or preferred alignment of primocrysts (commonly of elongate plagioclase crystals) (Irvine et al 1998).

Remarkably interesting features supposedly formed by a repeated deposition of density currents from nearby walls can be found in Skaergaard intrusion, predominantly in the Upper Border zone A. These synformal, modally graded layered structures of a crescentic shape are known as troughs (Fig. 10).

Each trough comprises a set of layers consisting of a mafic base and a plagioclase rich top, and has a sharply defined contact with both underlying and overlaying massive gabbro. Some of these structures can be traced as far as 100 meters throughout the stratigraphy. Troughs were firstly described as the result of repetitive deposition from crystal-rich currents by Wager and Deer (1939). These authors substantiated the idea by pointing out the sedimentary features the cumulates preserve, such as the preferred orientation of plagioclase grains parallel to the modal layering or lateral migration of layers. The hypothesis was later challenged by McBirney and Noyes (1979) as they attributed the formation of troughs to in-situ growth. However, recent complex studies of the fabrics and microstructures done by Holness et al (2017b) and Vukmanovic et al (2018) support the sedimentary origin of these unusual layered structures.

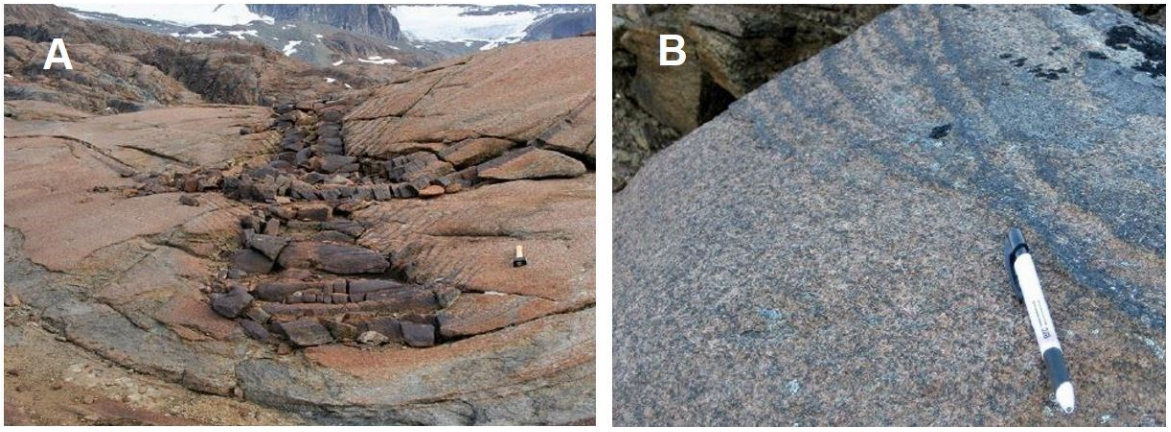


Figure 10 – **a)** A field photograph of the Trough I in the Upper Border zone A. **b)** A close-up shot of one of the loose blocks surrounding Trough O also in UBa, showing evidence of layer truncation caused by migration of the trough axis. From Vukmanovic et al (2018).

IV.1.4. Magma chamber replenishment

Layered intrusions are oftentimes replenished by a pulse of fresh magma numerous times throughout their solidification history. In fact, multiple recharge events are presumed in a majority of cases as intrusions created by a single pulse of magma appear to be a rare occurrence, especially amongst bigger and stratigraphically complicated bodies. Repeated replenishing is often associated with the formation of cyclic layering (Irvine and Smith 1967), but it can also contribute to the formation of cumulates by bringing a load of new phenocrysts.

In case the new magma pulse began to cool down during its ascend through the feeder conduits, it is likely that it will not be crystal-free upon entering the chamber. Carrying a load of phenocrysts in suspension, the fresh magma enters the intrusion and spreads out across the floor as a basal flow. If the crystal cargo is sufficiently small, it may allow the flow to cover most or all of the chamber floor. As the new pulse settles in, the crystals carried in suspension will begin to settle and form a new cumulate unit (Naslund and McBirney 1996, Latypov and Chistyakova 2020).

Cumulates formed by a deposition of crystals from a flowing (and possibly convecting) basal flow may display strong lamination of elongated grains parallel to layering (Holness 2020). As the magma can possibly carry several mineral types each with a variety of grain sizes, modal and grain-size layering is also likely to develop. However, given that similar features can result from any of the previously mentioned processes and/or their concurrent operation, it is a matter of interpretation to decide whether a replenishment by a crystal bearing magma played an important role in a specific case. Recently, this process has been used as a possible explanation for the origin of plagioclase-chromite adcumulates from the Critical Zone of the Bushveld complex (Latypov and Chistyakova 2020).

IV.2. Cumulates formed by in-situ nucleation and growth

Even though the idea of crystal redistribution offers a plausible explanation for many of the structures observed in layered intrusions, there is just as many features accompanying them which cannot be easily interpreted as resulting from any of the previously mentioned processes. The most prominent ones were first called out by Campbell (1978) in his work regarding the cumulate rocks of Jimberlana intrusion, followed shortly by McBirney and Noyes (1979) with their investigation of Skaergaard layering. Based on their observations and experimental work, they made a relevant presumption than in-situ nucleation and growth might be equally, if not more important process to cumulate formation than was formerly assumed.

In-situ formed cumulates can form anywhere throughout the magma chamber regardless of the inclination of the surface (on the temporary floor, walls and on the roof), growing inwards toward the centre of the intrusion. The nucleation mechanism is presumed to be exclusively heterogenous, and growing crystals on top of the crystal pile are supplied with new material by a combination of convection and diffusion. This is essentially a process similar to the adcumulus growth (Campbell 1978, McBirney and Noyes 1979). However, said similarity does not imply that in-situ formed cumulates can only be adcumulates. Whether the rock will have an orthocumulate or adcumulate texture is ruled by conditions at the interface between the crystal mush and the surrounding liquid. If the growth of grains stops at the moment the crystals start coming into direct contact and forming a connected network, the resulting rock will enclose large amounts of interstitial liquid and thus become an orthocumulate. If however the primocrysts keep the chemical communication with the bulk magma long enough and the growth continues, pushing out the remaining liquid, it will result in an adcumulate texture.

Overall, in-situ cumulates are difficult to define in terms of their usual attributes, as there is no general agreement on which textural aspects of cumulates can and cannot be formed by in-situ crystallization. In an ideal case, it is quite intuitive that without the influence of transport and deposition processes, in-situ formed cumulates should not preserve any types of sedimentary-like features. There should also typically be a wide range of grain sizes amongst primocrysts as a result of absent hydrodynamical

sorting, without any shape preferred orientation of primocrysts (Holness 2020). However, some rocks commonly interpreted as in-situ cumulates deviate far from such ideal case. For instance, dendritic crystals in crescumulates exhibit strong shape preferred orientation. Moreover, it is important to clarify that formation in-situ oftentimes does not equal an unlayered rock, as there are many examples of cumulate rocks exhibiting prominent layering of diverse types which have been interpreted as formed in-situ (McBirney and Noyes 1979, Boudreau 2011, Namur et al 2015). Layering in such cumulates develops differently by processes of more chemical than mechanical nature, some of which will be discusses further in this chapter.

IV.2.1. Cumulate rocks formed on steep surfaces

Cumulate layers in mafic intrusions are commonly subhorizontal or gently dipping, stacked on top of each other throughout the height of the chamber. Such positioning does not oppose to the formation by crystal settling, density currents etc. However, the same cannot be said about layering found on the walls of the very same intrusions, which can be in some cases nearly vertical. Jimberlana intrusion in Western Australia is a convenient example; it is a funnel-shaped body, uncommonly narrow when compared to other layered intrusions such as Muskox or Bushveld, with horizontal layering in the centre and steeply dipping layers at the margins. There is no abrupt angle or sharp discontinuity between the central and marginal layering, instead the layers are smoothly interconnected. There also does not seem to be any notable textural difference between the rocks of one layered unit throughout its length, regardless of the angle of inclination (Campbell 1978, Campbell 1987).

Such observations are obviously inconsistent with many aspects of the cumulate theory and operation of crystal redistribution processes on a large scale, instead are much more easily reconcilable with formation in-situ. The reason why this was not considered as much of a struggle when Wager and his colleagues formulated the cumulate theory is the fact that majority of research has been based on Skaergaard intrusion (Wager and Deer 1939, Wager et al 1960), which is different from Jimberlana in many aspects. In this context, it is mainly the absence of angular discontinuity between the marginal zone and central layering in Jimberlana, which on the contrary is very prominent in Skaergaard, especially on its eastern and northern margin (Fig. 11) (Wager and Brown 1968).

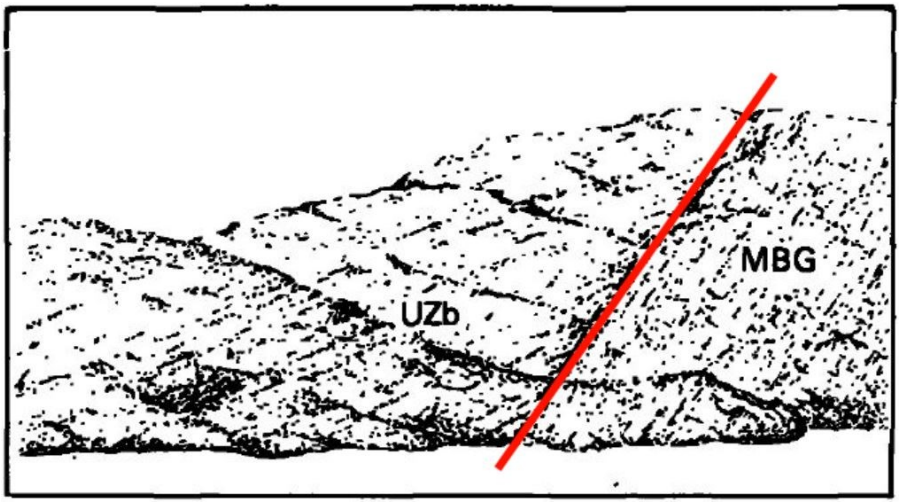


Figure 11 – An illustration of the relations between the Marginal border group and Upper zone B in the northern part of Skaergaard intrusion, showing the sharp angular change between both series. Modified after McBirney and Noyes (1979).

Skaergaard main layered series thus at the time did not directly contradict with the idea that crystal settling on a horizontal surface must be the main formation mechanism for cumulates in layered intrusions. Wager and Brown (1968) did acknowledge that the marginal series on the walls and against the roof required another mechanism to operate besides settling and presented a couple of possible processes, including attachment of crystals on the walls while being carried downward by the convective currents, or a process essentially similar to what we know today as in-situ nucleation and growth of cumulates (Wager 1961). However, the importance of the latter was obviously greatly underestimated at the time as it was associated with chilled margins only.

As represented by the cumulates of Jimberlana intrusion, rocks found on the steep walls can have similar structural and textural attributes as the subhorizontal cumulates in the centre, up to a point where two rock samples would be indistinguishable when compared side by side. (Campbell 1987). There are mostly no distinct properties to them other than the steep angle of layering and occasional dendritic growth of primocrysts. Primocrysts can have shape preferred orientation perpendicular or parallel to layering or be completely order less (Holness 2020).

An interesting feature occasionally found in the marginal zones on steep walls of layered intrusions is a specific type of cumulate rocks, known as crescumulates. These rocks are distinguished by large dendritic crystals, most often olivines, growing perpendicularly to the contact with the surrounding rock and to the layering, following the direction of heat loss (Namur et al 2015, Holness 2020). Crescumulates occur almost exclusively near the margins of the chamber as they require significant undercooling in order to form, and thus are rarely found elsewhere amongst other cumulates in any main layered series. In Skaergaard, such rocks occur in the Border Zone approximately 10 meters away from the contact with country rock (Wager and Brown 1968, McBirney and Noyes 1979, Irvine 1982). The rare example of ‘displaced’ crescumulates can be found in the Rhum intrusion, far from the edge

where they formed presumably by growing upwards into a supercooled liquid (Irvine 1982). Either way, crescumulates are generally interpreted as cumulates formed by growing in place and no process of crystal redistribution is being attributed to their existence. As such, they represent a rare consensus in the field of cumulate research.

IV.2.2. The plagioclase problem

Plagioclase with its positive buoyancy in ferrobasaltic liquids has been a notorious troublemaker in a history of cumulate research, especially when discussing the primocrysts in Skaergaard intrusion. Using chemical compositions estimated by Wager and Brown (1968), Bottinga and Weill (1970) calculated that the Skaergaard parental liquid could have been as much as 0.04 g/cm^3 heavier that the plagioclase primocrysts that crystallized from it. This would imply that the density contrast between the magma and plagioclase was sufficient for the grains to float instead of sink. Their calculations have later been repeatedly tested by other researchers, mostly with similar conclusions. For example, based on their series of experiments in centrifuge furnace, Campbell et al (1978) concluded that plagioclase's tendency to float in Skaergaard magma could be even greater than what the original calculations indicated.

The problem of floating plagioclase of course does not affect all layered intrusions, solely those with high enough density of their parental liquid(s). However, with Skaergaard being one of these intrusions while being in the centre of attention of cumulate research sparked doubts about the legitimacy of Wager's interpretations (McBirney and Noyes 1979). Even though the plagioclase grains should have floated in the Skaergaard liquid, there is no significant accumulation of plagioclase preserved under the roof of the intrusion as would be expected (Naslund 1984). Instead, the plagioclase primocrysts can be found in cumulates all across the main layered series, beginning with its lowest segment known as Lower Zone A (McBirney and Hunter 1995, McBirney 1996). These observations seem to work in favor of in-situ nucleation and growth at the temporary floor of at least some of the plagioclase cumulates in Skaergaard intrusion, and consequently in other intrusions with similar compositions. Pre-existing crystals on the floor of the chamber provide opportunity for heterogenous nucleation of plagioclase, making its buoyancy irrelevant.

Several other hypotheses regarding the occurrence of plagioclase on the floor of Skaergaard assume that regardless of the significant density contrast, plagioclase cumulates still could have formed as a result of gravitational settling, but only if the grains settled in clusters or chains involving other mineral phases nucleating in the liquid such as olivine or pyroxene (Campbell 1978). The weight of the cluster would then be sufficient for plagioclase to sink and subsequently form cumulates on the floor. However, these hypotheses are inconsistent with two widely occurring types or cumulates: With those in which the primocrysts have strong preferred orientation and preserve no evidence of chain settling,

and with monomineralic plagioclase layers in ferrobasaltic intrusions. Overall, formation in-situ appears to be more universally applicable solution for seemingly misplaced plagioclase primocrysts.

IV.2.3. Fine-scale rhythmic layering

Rhythmic layering on small scales of millimetres to centimetres, found in a majority of layered intrusions throughout their stratification. The most well-known example of prominent fine-scale layering, around which most of the research on the topic is based, can be found in norites from the Stillwater complex and is defined by alternations of pyroxene and plagioclase rich layers each only a few centimetres thick (Jackson 1961). Such feature is a very problematic attribute if we assume that layering is produced solely by sedimentary-like processes. In the past, the formation of rhythmically layered cumulates has been attributed to density currents, which periodically swept across the floor and each time deposited a modally sorted cargo of crystal (Irvine 1980a); an idea which is still valid in some cases, but obviously troublesome on the microrhythmic scale. There is now a number of hypotheses on the formation of this type of layering (e.g., Brandeis et al 1984, McBirney and Noyes 1979), most of which do not consider crystal redistribution to be of major importance and instead associate the formation of microrhythmically layered cumulates exclusively with in-situ nucleation and growth.

One of the most widespread hypotheses attributes the formation of rhythmic layering to periodic nucleation of given mineral phases (McBirney and Noyes, 1979). Periodic nucleation is explained as follows: A layer of liquid in the proximity to a region of solidification may become supersaturated with one or more mineral phases, in this case the phase ‘A’. As mineral ‘A’ begins to nucleate and grow, the components necessary for its growth will begin to diffuse toward the growing crystals and create a zone depleted in said elements in the surrounding liquid. The zone of depletion progressively spreads, however, its boundary will move slower with time and will be eventually overrun by the thermal gradient in the crystal mush which advances at a virtually constant rate. As the zone of cooling outruns the boundary of depletion, a new burst of nucleation of the ‘A’ phase will occur. Between the two episodes of ‘A’ nucleation, the depleted liquid can become supersaturated in phase ‘B’, and the whole repetitive process results in a formation of rhythmic layering defined by layers dominated by ‘A’ and ‘B’ respectively. Many of the other hypotheses, for example based on different rates of diffusion of two or more components (Boudreau 2011), work on a similar principle.

The periodic saturation model introduced by McBirney and Noyes (1979) has been used to further illustrate that not all seemingly sedimentary-like features in layered intrusions should automatically be taken as such. Moreover, the fact that microrhythmic layering of various extend can be found not only on the chilled margins but in the mail layered series of most intrusions as well, further supports their view that in-situ nucleation and growth of cumulates might possibly be the dominant formation process overall.

There are naturally various other opinions on this topic. In a more recent work centred around Stillwater layering, Boudreau (2011) acknowledges that the model of periodic nucleation is undoubtedly valid but expresses doubts about its effectivity on a large scale. Instead, Boudreau assumes that fine-scale layering in fact does not form during the accumulation of crystals and thus does not depend on whether the crystals nucleated and grew in place or accumulated in other way, but rather forms during the later stage of solidification as the crystals age and texture further develops – the stage of postcumulus processes.

V. Formation of cumulates II: Postcumulus processes

Following the accumulation of primocrysts, the formation of cumulates enters a second stage where chemical and mechanical processes in the crystal mush shape the texture of the rock into its final form. The so-called 'postcumulus processes' mainly include the crystallization of the interstitial minerals from the trapped melt and the adcumulus growth, which can together be view as processes having a first-order effect on the developing rock (Irvine 1982, Tait et al 1984). Adcumulus growth and crystallization of intercumulus liquid act as competing mechanisms to each other and the predominant process predetermines the character of the forming texture, resulting in either orthocumulate, mesocumulate, or adcumulate texture (Hunter 1996). Postcumulus processes also include other potential influences which may alternate the texture further, for example compaction (Sparks et al 1985, Holness et al 2017) and metasomatic replacement (McBirney and Sonnenthal 1990).

In order to understand the mechanics of postcumulus processes, especially the formation of adcumulate texture, both theoretical and experimental studies chiefly focus on the fluid dynamics in the crystal mush, such as melt circulation, compositional convection etc. (e.g., Tait et al 1984) and on the influence of compaction processes in the deeper levels of the mush (e.g., McKenzie 2011, Holness et al 2017), both in respect to the field observations, chemistry of the rock and microstructural record. The main challenge, particularly when discussing the early stages of textural development, lies in the fact that the rocks themself might present a misleading source of information. The original record of processes operating in the solidifying mush might have been overprinted several times by later alterations of both texture and geochemical signatures (Boudreau 2011).

V.1. Initial porosity

Two types of porosities can be distinguished when describing cumulate rocks; The initial porosity, aka the amount of interstitial liquid in a crystal mush immediately after the accumulation of primocrysts, and the residual porosity, which is defined as the modal amount of interstitial phases and postcumulus overgrowths in a solidified rock (Irvine 1982). Only the residual porosity can be directly measured from field samples, providing that the rock has not been considerably altered by subsequent metasomatic or metamorphic recrystallization (Morse 1979b). In case of initial porosity, the number can only be estimated by using theoretical and experimental methods unless the rock in questions has a perfect orthocumulate texture (Irvine 1982).

Initial porosity of a cumulate and our ability to estimate its value largely depends on the dominant formation mechanism (Campbell 1987, Hunter 1996). In a cumulate rock formed mostly by crystal redistribution, the porosity is defined by packing of accumulated primocrysts, which is controlled by several factors such as grain shapes, the range of grain sizes (size sorting) and preferred orientation of

non-spherical crystals (Holness et al 2020). A number of theoretical studies and analogue experiments have been performed in the past to quantify the porosity for a variety of cases (in terms of variable grain sorting and shapes). For instance, while formulating the original cumulate theory, Wager et al (1960) performed an experiment using cuboid-shaped aluminium tablets in an aquatic suspension to simulate the packing of primocrysts in a plagioclase orthocumulate in Skaergaard Lower Zone A, resulting in 40 to 60% initial porosities. Theoretical modelling based on spatial distribution of grains (e.g., Jerram et al 1996) provided comparable results. A similar number, approximately 40-50% porosity, was obtained by measuring porosities in well-sorted sandstones containing mostly rounded grains (Atkins and McBride 1992).

Cumulates formed predominantly by in-situ nucleation and growth require an entirely different approach, given by the fact that the grains grow in place and so the initial assemblage of primocrysts in the crystal mush is not controlled by their mechanical arrangement. Instead, the outlook of crystal framework is controlled by nucleation density and the degree of undercooling (Hunter 1996). As such, the estimates of 40-60% initial porosity for redepositional cumulates are inapplicable in this case and there is essentially no evidence supporting the assumption that the initial porosities of in-situ cumulates had to be as high (Campbell 1987, Boudreau 2011). In fact, the term 'initial porosity' might even be inappropriate for this type of cumulate formation (Hunter 1996).

V.2. Adcumulus growth

Adcumulates are defined as cumulate rocks formed exclusively of unzoned primocrysts crystallized from the bulk magma, with very little or none interstitial material left in between the primary crystals (Wager et al 1960) The limiting percentage of intercumulus phases for adcumulates is around 5 %, however, in literature the term 'adcumulate texture' is often being used much more loosely. With so little trapped liquid remaining in between the primocrysts, adcumulates represent the most extreme end member of the cumulus terminology resulting from (almost) complete expulsion of the pore liquid after the accumulation of primocrysts. The processes driving the growth of primocrysts have collectively been termed as 'adcumulus growth' (Wager et al 1963, Irvine 1982).

Adcumulate rocks and the processes of their formation currently present the overall most intensely studied topic in the field of cumulate research. The first hypotheses formulated by Hess (1960) and Wager (1963) assumed that the adcumulus growth is enabled by diffusion of unwanted components away from the trapped liquid into the bulk magma. The less interstitial phases left, the more effective was the diffusion between the mush and the main magma. Wager concluded that in order for this process to be efficient, adcumulus growth should take place in shallow levels of a relatively thin crystal mush (as to keep an efficient chemical communication with the bulk magma and ensure the removal of latent heat of crystallization by convective currents) or/and where the rate of accumulation is slow. Therefore, in case the conditions are reversed (aka thick crystal mush or fast accumulation of

primocrysts), the adcumulus growth should be severely limited which favours the formation of orthocumulates.

At present, adcumulus growth controlled solely by diffusion is considered merely as one of the possible mechanisms, rather than the main driving force behind adcumulate formation. Modern studies mostly assume that the required conditions favouring adcumulus growth are likely to be the exact opposite of what Wager described, given that the processes we now consider as important factors, namely compaction and compositional convection, are considerably more effective in deeper parts of the crystal mush (Holness et al 2017).

Overall, the suggested mechanisms enabling adcumulus growth can be split into two broad groups: The first group of researchers, including aforementioned Wager and Hess, focuses on processes of mostly chemical nature. These include the interplay between the nucleation and growth during in-situ formation (Campbell 1978, Morse 1986), the diffusion of components from the bulk magma (Wager et al 1960, Hess 1960, Wager 1963), compositional convection (Tait et al 1984, Tail and Jaupart 1992) and the subsequent textural equilibration (Cabane et al 2005, Higgins 2011). The second group incorporates processes leading to mechanical expulsion of the interstitial liquid from the crystal mush, such as mechanical and viscous compaction (e.g., Sparks et al 1985, Tegner et al 2009, Holness et al 2017), and penetrative shearing (Bédard 2015). All of the processes mentioned above are considered relevant, with each playing a predominant role under specific conditions. However, at present there are no clear guidelines or established criteria on how to recognize the influence of each process from field samples or chemical analysis, and so the current efforts are mostly directed towards deciphering the imprint of each process in the microstructural record (Holness et al 2017).

V.2.1. Thermal convection and diffusion at the magma-mush interface

When discussing adcumulate formation in shallow parts of the crystal mush, especially if driven by diffusion, large scale convective currents are considered as another important factor permitting the adcumulus growth apart from slow accumulation (Wager 1963). As the convection currents continually sweep across the temporary magma chamber floor, the flow carries away any rejected solute produced by crystals growing within the crystal mush and rising up due to its generally lower density. Morse (1968) estimated the velocities of convective currents in a large mafic intrusion (km/day to km/year) and the flux of rejected liquid from the mush (driven by the rate of crystal growth, approximately in the order of cm/year). Given the five to seven orders of magnitude difference between both velocities, Morse (1968) concluded that the rejected solute will not stagnate above the interface for any significant amount of time before it is carried away. The convective currents thus continually bring fresh magma for the growing crystals near or at the interface, while carrying away the rejected liquid and the latent heat, promoting an effective adcumulus growth.

V.2.2. Compositional convection

Compositional convection is considered as a crucial mechanism for adcumulus growth, especially in the early stages following the accumulation when permeability of the mush is still relatively high. Unlike the diffusion mechanism relying on thermal convection at the interface, compositional convection is not restricted to the shallow levels of the mush and can operate throughout the whole volume of the cumulate pile if the conditions favour it (Tait et al 1994, Hunter 1996).

The process is based on a vertical exchange of liquids according to their density differences. When mafic minerals crystallize and grow, they become surrounded by a thin film of liquid depleted in mafic components progressively becoming lighter than its original composition. The lighter melt is then able to convect upward and away from its original place. The lighter liquid is replaced by a downward flow of denser, more primitive melt from above and provide new material for adcumulus growth. The same principle applies to the growth of plagioclase, where the produced melt becomes denser than its surroundings and tend to move downwards through the mush while being simultaneously replaced by lighter melt from beneath or denser melt from above. (Sparks and Huppert 1984, Tait et al 1984). However, it is generally assumed that compositional convection is much more effective in assisting the adcumulus growth in mafic cumulates, where lighter liquid is produced (Sparks et al 1985). The possible effect of magma chamber replenishment is also often considered, especially when a dense primitive magma is emplaced on the floor of the chamber above the cumulate pile. Since the newly emplaced magma is likely denser than the interstitial liquid inside the mush, the new influx will begin to sink the lighter melt will rises (Huppert and Sparks 1980, Sparks et al 1985).

Importantly, since the efficiency of compositional convection drops significantly with the decreasing permeability, this process does not aid the adcumulus growth any further once the porosity of the pile lowers (Tait et al 1984). In the low porosity conditions, the melt will either get trapped, or possibly concentrate into flow channels and move through paths of least resistance (Tait and Jaupart 1992, Hersum et al 2005). As such, compositional convection alone likely cannot be responsible for the formation of adcumulates *sensus stricto* (aka less than 5 % of interstitial material) in a majority of cases and is expected to produce mostly mesocumulates (Tait et al 1984, Sparks et al 1985). Moreover, high permeability is only one amongst other conditions needed for an efficient compositional convection. For example, experimental studies have demonstrated that adcumulus growth driven by compositional convection is realistic only if the convective velocity of the interstitial melt is greater than the solidification velocity. In turn, the smaller, quickly cooled intrusions are not a favourable environment for this process (Sparks et al 1985). Nevertheless, compositional convection is a plausible mechanism assisting the adcumulus growth during the high-porosity stage of the crystal mush development, especially for mafic cumulates in larger chambers.

V.3. Compaction

Compaction is essentially a result of imbalance between hydrostatic and lithostatic pressure inside a cumulate pile, which is forcing the interstitial liquid to move upward through the mush due to being significantly lighter than the solid primocrysts. As such, it effectively lowers the porosity of a forming cumulate by removing the remaining trapped melt, which may result in an adcumulate texture (Irvine 1980, Sparks et al 1985, Tegner et al 2009). However, it is important to note that although compaction is generally accepted as a significant factor in adcumulate formation, it is not included amongst processes causing adcumulus growth as it does not directly promote the growth of primocrysts. In an extreme case, the compaction might thus be able to produce a perfect monomineralic adcumulate without any adcumulus growth taking place during its formation (Sparks et al 1985). In a more general perspective, compaction is thought to operate alongside compositional convection and other processes responsible for adcumulus growth, but it is nevertheless viewed as a separate set of mechanisms.

When discussing adcumulate formation, it is important to highlight the distinction between mechanical and viscous compaction. Mechanical compaction corresponds to an overall reduction of volume of a porous environment (in our case the crystal mush) in its early stages of development, caused by mechanical reorganization of solid particles (Holness et al 2017). Ability of the environment to resist mechanical compaction is given by the properties of grains and grain boundaries, and by its location specifics. This process it thought to operate regularly in crystal mushes, both locally and on a large scale. Reorganization of primocrysts can be initiated for example by a flow of magma by density currents on the magma-mush interface, or by a flow of the mush itself via slumping and sliding (Hunter 1996). Field evidence illustrating the significance of this process can be found for example in a 300 meters thick Palisades sill in New York, where plagioclase cumulates bear evidence of mechanical compaction due to the collapse of the initial crystal framework by approximately 10 % (Gray et al 2003).

However, in the context of adcumulates, the word ‘compaction’ usually refers to viscous compaction unless specifically stated otherwise. This type of compaction reduces the mush porosity by viscous grain-scale deformation/flow of the crystal framework, driven by processes such as dissolution-reprecipitation, dislocation creep, diffusion creep, etc, which are capable of lowering the porosity much further than the maximum packing density (Sparks et al 1985). The mechanism thus creates a two-phase flow, with an upward movement of the interstitial liquid and compaction of the solid framework. As a gravitationally driven process, viscous compaction requires a thick crystal mush and takes place in its lower levels, where the pressure created by the mass of overlying primocrysts is sufficient (Holness et al 2009, Holness et al 2017).

This concept was first explored (in connection with adcumulates) by Irvine (1980), who described a series of cyclic units in Muskox intrusion with seemingly disordered discontinuities in chemical

compositions in respect to modal mineralogy of each layer. He argued that the shifts in mineral compositions throughout the units were caused by an upward flow of melt thought the mush, which has been expulsed from its interstitial spaces by viscous compaction. The process was later theoretically analysed in terms of rheological properties of olivine-dominated crystal framework (e.g., McKenzie 1984, Sparks et al 1985) and applied to a broader set of compositions by McKenzie (2011). Present research including the latest review on the topic by Holness et al (2017) shifts its attention towards the microstructural record and on how to recognize an influence of each possible compaction mechanism in a thin section.

V.3.1. Dissolution-reprecipitation

Dissolution-reprecipitation (also known as pressure-solution) is a diffusive transport of mass in a solution to regions with lower stresses. This process is essentially similar to diffusion creep, which is always associated with grain boundary sliding accommodating the spatial changes in grains (Holness et al 2017). The transport of mass starts at the regions with high thermodynamic potential, aka the points with higher deviatoric stress, which in this case are the point contacts between individual grains and their strongly curved edges. At these points, the grains dissolve and the material is transported to where the deviatoric stress and the surface energy contributions are lower (regions with lower potential), that is, onto the flat faces of the grains (Fig. 12). There are three possible pathways the diffusion may choose, depending on current local conditions, with each possibility having different rates and activation energies: Transport along the grain boundaries (the Coble creep), through the grain volume (the Nabarro-Herring creep), and through the pore spaces filled with trapped melt (the MED creep). The MED creep dominates while the interstitial melt fraction is still relatively high, and is then outweighed by the grain boundary diffusion as the compaction progresses (Cooper and Kohlstedt 1986, Hunter 1996).

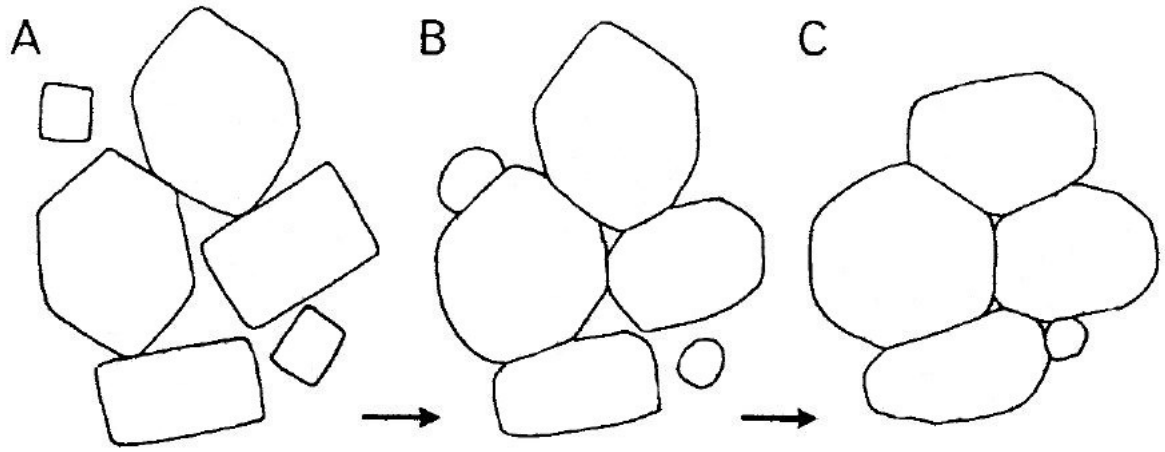


Figure 12 *– Progressive changes in grain shape and reduction of porosity by dissolution-reprecipitation. The dissolution concentrates at the sharp edges and contacts of the grains, and the lost material reprecipitates again at the low-curvature points. From Hunter (1996).*

The transfer of mass thus acts in a way as to minimize the surface energies and deviatoric stress, meaning that the texture resulting from dissolution-reprecipitation should be mature and equilibrated (Hunter 1987). This aspect however cannot serve as a relatable lead, given that textural equilibration is achieved in many other ways.

Admittedly, textures created by diffusion controlled viscous compaction are commonly hard to identify. Besides the observations of relationships between preferred orientation of crystals, which may point to diffusive deformation taking place, another important factor is a compositional zoning of crystals. All processes driving viscous compaction, and dissolution-reprecipitation especially, are able to partly obscure or even completely overwrite the initial compositional zonation formed during the early stages of crystallization. As such, this process causes irregularities such as a local loss of planar growth faces or a core not corresponding to its original shape, which may possibly serve as an important hint in an attempt to identify the diffusion-based compaction mechanisms (Holness et al 2017).

V.3.2. Dislocation creep

Dislocation creep is a compaction mechanism characterized by gliding and movement of dislocations within the grains, which occurs in response to differential stress. Grains are divided into subgrains by the dislocations and deform along the resulting slip systems, resulting in re-orientation of the crystal framework (Fig. 13). In addition, dislocation creep is accompanied by recrystallization which can erase dislocations or transform them into grain boundaries while new dislocations proceed to form elsewhere (Hunter 1996, Meuer and Boudreau 1998, Holness et al 2017).

Dislocation creep always results in both crystal and shape preferred orientations, where CPO forms according to the dominant slip systems and SPO accommodates the shape and volume changes caused by the deformation (Tegner et al 2009), Holness et al 2017). Sometimes, more apparent signs of deformation such as bending of crystals may also be present (Hunter 1996). A total overprint of these fabrics is not threatened by the recrystallization if the temperature drops relatively quickly, however, in case the temperature is persistently high, static recrystallization will continue and erase all evidence of shape preferred orientation (and most of CPO as well), the texture will progressively become granular and approach textural equilibrium (Higgins 2015). Signs of static recrystallization can be found mostly locally in solidified shallow magma chambers (mostly in monomineralic regions) but are expected to be of greater importance with increasing depth and size of the magma body (Hunter 1996).

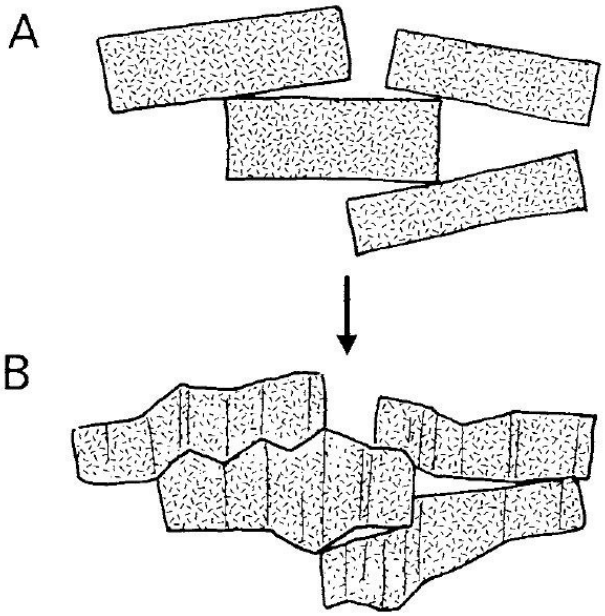


Figure 13 *– The effects of compaction by dislocation creep on the shape of tabular plagioclase grains. The pressure causes the formation of dislocations (illustrated in the stage B as lines perpendicular to the C axis of the grain) and bending of the crystals. From Hunter (1996).*

V.4. Textural equilibration

The entire topic of textural equilibration is surrounded by a lively discussion regarding the credibility of microstructures and processes they preserve (Boudreau 2011, Holness et al 2018). Texture of a cumulate rock is shaped during several stages, each capable of completely overprinting the previous phase. The initial textures produced in crystal-poor magma, also termed as 'kinetic' or 'kinetically-controlled', are formed solely by nucleation and growth of crystals. These textures are commonly preserved in volcanic rocks, but much less in coarse-grained plutons where they are likely to be reshaped by subsequent processes (Higgins 2015). The kinetic texture is further modified under the influence of mechanical processes, such as redistribution or compaction, which operate while the melt is still present in the system (Higgins 2015, Holness et al 2017). Both the initial and mechanical textures are not equilibrated, meaning they have high total surface energies. If a sufficiently high temperature is maintained for a prolonged period of time, the texture enters an equilibration stage.

The vast majority of rocks in layered mafic intrusions have been equilibrated to a variable extend, and under a range of thermal conditions (Hunter 1996). Textural equilibration is defined as a reduction of total surface energy by changes in area and geometry of individual grains. It occurs both in presence of melt and in sub-solidus, although the first significantly enhances its effectivity (Hunter 1987, 1996). The textural changes begin at grain boundaries, firstly by adjusting the dihedral angles to more convenient values (less than 60°, Holness 2010) and then proceeding to even out the curvature of the grain and change its shape towards more equant proportions (Higgins 2011, Holness et al 2006). As an example, in a study of a layered troctolite body, Higgins (1998) stated that while unequilibrated

plagioclase crystals enclosed in an olivine oikocryst have an aspect ratio of 1:3:3, the value for more equilibrated grains in a surrounding matrix is approximately 1:1.5:1.5. Following these modifications, surface energy is further reduced by coarsening.

Also known as Ostwald ripening, the coarsening process is based on dissolution of small grains and transfer of their mass in favour of the growth of larger crystals, naturally resulting in a larger mean grains size and lower mean number of grains per unit volume (Park and Hanson 1999, Higgins 2011). Larger crystals have lower mean curvature of their shape, meaning that the total boundary area per unit volume is reduced and so is the surface energy. However, for Ostwald ripening to have a significant impact on the grain size distribution, very low degrees of undercooling are needed (likely under 1 °C) otherwise kinetic crystallization overruns its driving force (Holness and Vernon 2015). Also, as the experimental studies have shown, the effect of coarsening dramatically lowers with increasing grain size, and overall, the process requires long periods of time to strongly alternate the texture. On a shorter time scale, Ostwald ripening might only be capable of reducing the number of small crystals in their recent post-nucleation stage (Higgins 2015).

Different types of minerals phases are affected by Ostwald ripening at different rates, favouring those with higher diffusion rates and higher solubility. The set of experiments performed by Cabane et al (2001, 2005) demonstrated a significant difference between the coarsening rate of quartz and plagioclase, which both showed a very slow response, and the contrasting fast rate of coarsening of olivine in a basaltic melt over much shorter time period. For plagioclase, this process proved effective only for the grain sizes below 20 μm, even over time scales of hundred thousand years. The overall importance of Ostwald ripening on textural development in layered intrusions is still a matter of discussion, but a majority of researchers agree upon the fact that this phenomenon might be enhanced rather sporadically, for example following a larger nucleation event accompanying an arrival of a new liquidus phase (Campbell 1987, Cabane et al 2005).

VI. Experimental investigation of cumulate formation

As our current technology does not allow us to directly observe processes inside magma chambers during their active lifetime, a significant part of petrological research relies on experimental data to alleviate this disadvantage. Formation of cumulates is no exception in this matter; Attempts to simulate the mechanisms which are assumed to operate in layered intrusion have always accompanied the theoretical studies, from early investigations of crystal settling and initial porosity (e.g., Wager et al 1960) to high-pressure experiments with intend to induce compaction of a cumulate layer (e.g., Schmidt et al 2012).

In this work, we aimed to observe a behaviour of two contrasting crystal suspensions over a period of time. In liquid-rich environment, crystals are expected to settle down to the floor of the experimental capsule and thus create a closely packed, size-sorted redistribution cumulate. If however the crystallinity of the environment is high, it is assumed to prevent the movement of crystals throughout the capsule and provide an opportunity to observe other processes beside settling, such as equilibration. To achieve the intended conditions, we prepared two separate mixtures of basalt glass containing 10 and 60 wt. % of olivine crystal fragments and heated them in graphite capsules to simulate crystal suspensions in a mafic intrusion. A total of 14 experiments in two series differing by the temperature conditions were performed on a 20-hour long time scale, and resulting samples underwent textural analysis in order to describe and quantify the gradual changes in the system.

VI.1. Methodology

VI.1.1. Starting materials

As a starting material, we used a mixture of synthetic haplobasaltic glass and natural olivine crystals of typical mantle compositions (~Fo_{90}). To simplify the phase relations, the glass composition was chosen from the haplobasaltic forsterite-diopside-anorthite system (Fig. 14) with weight proportions of components corresponding to $Fo_{20}Di_{45}An_{35}$. This composition is located safely within the olivine saturation field with expected liquidus temperature of ~1400 °C and was chosen to resemble the composition of natural olivine normative basaltoids. Moreover, this composition prevents crystallization of other phases than olivine by keeping a reasonable distance from the Fo-Di cotectic and Fo-An-Di eutectic. The starting glass was obtained by melting a homogenized mixture of pure (99.8 wt. %) MgO, Al_2O_3, SiO_2 oxides and $CaCO_3$ carbonate at 1450 °C in a platinum crucible for 12 hours and by subsequent grinding and homogenisation in an agate mortar (Tab. 2). The resulting fine powder provided a base for the glass-olivine mixtures.

Olivine crystals were collected from the mantle xenolithes in Smrčí gravel quarry (Fediuk 1971) and then hand crushed to smaller fragments ranging from ~1 to ~110 μm in size, with a median size of

2.9 µm. Only fresh and clear crystals with typical olivine colour were used while the grains exhibiting colour or transparency variations were discarded.

Table 2 – Proportions of MgO, Al_2O_3, SiO_2 oxides and $CaCO_3$ carbonate in the starting material (wt. %).

Oxide / Carbonate	wt. %
SiO_2	42.4
Al_2O_3	11.2
$CaCO_3$	29.1
MgO	17.3

In order to observe the behaviour of both high and low crystallinity suspensions, we prepared two separate glass-olivine mixtures – one containing 10 wt. % (corresponds to 9.5 vol. %) of olivine crystals as an ideal proportion permitting crystal redistribution, and one with 60 wt. % (which corresponds to 58.6 vol. %) to inhibit the movement of grains inside the capsule. After mixing in corresponding proportions, the glass-olivine mixtures were thoroughly homogenized in ethanol. Both series of experiments altogether required approx. 10 g of synthetic haplobasaltic glass.

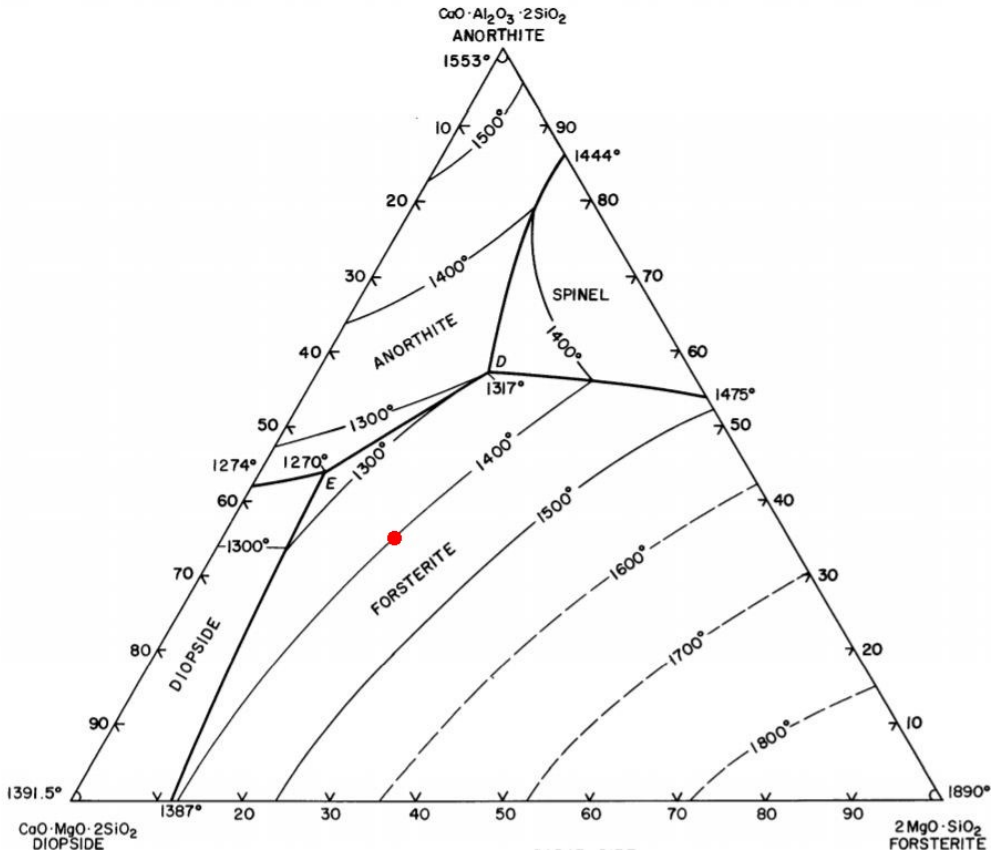


Figure 14 *– Liquidus phase relations in the haplobasaltic forsterite-diopside-anorthite system. The red dot corresponds to chosen composition of the synthetic basaltic glass. For the original diagram, see Presnall et al 1978.*

VI.1.2. Experimental techniques

As experimental capsules, we used 1x1x1 cm graphite cubes with hand drilled cylinder-shaped cavity, approximately 8 mm deep and 5.5 mm in diameter. The cavity was filled up to the top with the glass-olivine mixture and then covered with 2 mm thick graphite cover. As graphite would not be able to withstand the required temperatures when exposed to normal atmosphere, the cubes were thoroughly covered in multiple layers of fireclay in larger forms to prevent them from igniting. Once dry, the capsules were removed from the silicone, checked for any possible cracks in the fireclay and placed into the furnace for their assigned time, ranging from 22 minutes to 20 hours. The capsules were always heated in pairs, unless one sample required correction. At the end of each run, capsules were quenched in air or cold water and removed from the fireclay shells (Fig. 15).

All experiments were performed in atmospheric pressure and in identical setups, but were separated into series A and B due to differences in temperature. Series A consisted of four runs, each including one 10 wt. % olivine (A10) and one 60 wt. % olivine sample (A60), performed at 1390 °C with durations ranging from 22 minutes to 20 hours (a total of 10 experiments). For series B, the temperature was lowered to 1350 °C and only two runs were performed for comparative purposes (4 experiments total).

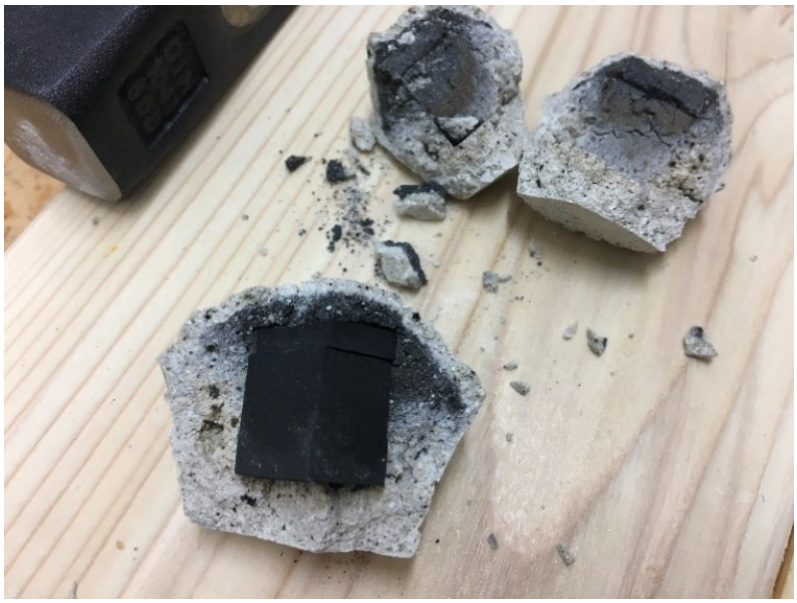


Figure 15 *– Process of removal of the protective fireclay shell from the graphite capsule, following a successful 4-hour long run of sample A10-240.*

VI.1.3. Analytical techniques

Following the quench, samples were cut out of the graphite cubes and mounted in epoxy, using a mounting press (SimpliMet 4000 by Buehler). Resulting mounts were hand ground by three types of sandpaper with gradually decreasing coarseness (grain coarseness number p500, p800 and p1200), polished by 5- and 1-micron corundum-water suspensions, and carbon coated in vacuum to prepare

them for analysis on an electron microscope. The scanning electron microscope Tescan Vega was used to obtain back-scattered electron (BSE) images of crystal-containing regions of the samples for textural analysis. During the BSE imaging, 15 kV accelerating voltage was employed and the brightness and contrast levels were adjusted to maximize the colour difference between the crystals and the glass. The resulting images were imported into a general-purpose image editing software, where each grain and the overall crystal-filled region were outlined in order to create a textural map for each sample (the darker quench overgrowths and dendritic crystals were not included).

For a complete textural analysis, an in-house software by Špillar and Dolejš (2014) was used to acquire data regarding the properties of the grains, including the grain sizes, the nearest neighbour distance, etc., as well as the total area of the analysed section. Finally, the R software (R Core Team 2020) was employed to calculate the quantitative textural parameters of interest for each sample: Crystallinity (Φ), clustering index (R, Jerram et al 1996) and mean crystal size (L, measured as the maximum Feret's diameter).

VI.2. Results

VI.2.1. Qualitative characterization

A total number of 14 experiments were successfully finished, from which 5 capsules did not withstand the experimental conditions of longer runs and needed to be remade. As expected, the samples with 10 wt. % olivine in both series have shown progressive settling of crystals, with all grains fully redistributed in under four hours (Fig. 16). A noteworthy detail is the gradual change in colour of the surrounding glass, from dark and barely translucent grey in shortest runs to completely clear and transparent in the longer experiments. Both A10 and B10 samples were generally compact, held together in one piece after the quench and contained only smaller bubbles, unlike their high-crystallinity counterparts.

Inside each of the samples containing 60 wt. % olivine, a large cavity was found taking up much of the space inside the capsule (Fig. 17). The cavities were remnants of large bubbles that pushed all the material towards the sides of the capsule, resulting in mostly O-shaped or U-shaped samples, which made them prone to breakage during grinding. All high-crystallinity samples were opaque, and the glass could not have been distinguished from the crystals with the naked eye.

Back scattered electron images provided a much more detailed look on the layout of crystals in the sample. As expected, the accumulated layer of crystals was found to be prominently size sorted in all 10 wt. % ol. samples, with most of the larger grains covering the floor of the capsule in under one hour. While still in suspension, crystals were observed to locally form small clusters (Fig. 18b). On the other hand, crystals in the high-crystallinity experiments were distributed randomly regardless of their shape

Table 3 – List of performed experiments, the experimental conditions and qualitative description of the samples from an optical microscope.

Name	T (°C)	Φ_0 (wt. %)	t	Quench	Description
A10-22	1390	10	22 min	Air	Crystals can be found throughout the whole volume of the sample, accompanied by a large number of acicular dendrites. Distribution of grain sizes is uneven, few of the largest grains form a pile at the bottom of the sample. The glass is translucent and has a sage green colour.
A60-22	1390	60	22 min	Air	The sample has a main cavity in its upper segment, and a couple of smaller holes and bubbles throughout the capsule. Crystals are distributed evenly throughout the intact regions in a manner reminiscent of stained glass.
A10-60	1390	10	1 h	Air	While the smallest grains remain in suspension in the upper part of the sample, the vast majority is concentrated at the bottom on a moderately size-sorted pile. The outlook of the glass is similar to the A10-22 sample.
A60-60	1390	60	1 h	Air	The sample is strongly disrupted by cracks and large bubbles, which left the compact crystal-filled regions only in upper parts around the perimeter of the capsule. Crystals are in close proximity, divided by grey translucent glass.
A10-240	1390	10	4 h	Air	Every grain from the suspension has accumulated at the bottom of the sample, while the upper parts are filled only with fully transparent light green glass. The accumulated crystal pile is noticeably size sorted. The interface of the cumulate is filled with skeletal, acicular dendrites growing into the overlying glass.
A60-240	1390	60	4 h	Air	A large cavity in the centre of the sample has confined the crystals at the bottom and edges of the capsule, resulting in a horseshoe shaped sample. The outlook of the crystal mush is similar to previous A60 samples.
A10-480	1390	10	8 h	Water	Similarly to the A10-240 sample, the crystal form a cumulate pile at the bottom of the capsule, while the overlying glass is crystal-free. The glass has a light grey colour and is transparent.
A10-1200	1390	10	~20 h	Water	The glass filling most of the sample is completely clear and transparent, with several large bubbles. The crystals are concentrated at the bottom; however, the accumulated pile is strongly disrupted by bubbles. No dendrites were observed.
A60-1200	1390	60	~20 h	Air	The U-shaped central cavity means that the crystal-filled regions are confined to the edges and upper part of the sample. The glass in between the crystals is transparent, and no dendritic crystals can be found.

B10-22	1350	10	22 min	Water	The sample has a small cavity in its uppermost part, but otherwise is compact. Most of the crystals are in suspension and there is no conspicuous accumulation of crystals on the floor. The suspension does not appear to be size sorted. Surrounding glass has a light grey colour and is translucent.
B60-22	1350	60	22 min	Water	Most of the crystals in the sample are concentrated on the bottom-right corner of the capsule, due to large bubbles taking up most of the space. Crystals are again in close proximity, and the pore spaces are filled with a dark grey glass.
B10-240	1350	10	4 h	Water	Similarly to the A10-240 sample, there is an accumulation of crystals at the bottom of the capsule. However, this sample appears to have been compromised during quenching, as the upper part of the cumulate was torn off by a large bubble and then pushed toward the roof of the sample. The transported piece of cumulate is made up of much smaller grains than the bottom part, meaning that the pile was likely size sorted prior to the damage.
B60-240	1350	60	4 h	Water	The crystal-filled regions of the sample make up most of its lower segment and encircle a large bubble around the perimeter near the roof. The crystals thus have a lot of space and appear to be less compacted than in the previous B60 sample, and the surrounding dark green glass is also easier to distinguish.

***List of variables**: T – temperature (°C), Φ_0 – initial crystallinity, aka olivine content in the mixture (wt. %), t – time, as duration of the experiment.*

and size and evenly filled out the available space (Fig. 17), proving that the chosen proportions of the starting material inhibited any significant movement inside the capsule. Contrary to our expectations, the chosen temperature difference between series A and B did not result in any major qualitative difference between the samples, and no signs of remelting have been found in the lower temperature runs.

A common attribute observed in every sample, regardless of its initial crystallinity or the quench method, was a prominent zonation of the grains. Each crystal, appearing as a light grey fragment, was surrounded by a darker rim of variable shape and thickness (Fig. 18a). The border between the original crystal and the rim was mostly sharply defined, but the transition seems to be much smoother in both B-240 samples when compared to series A equivalent. The rim appeared darker on the back scattered images due to compositional difference, as it could not have drained any iron from the synthetic glass and thus grew as pure forsterite. In the case of air-quenched 10 wt. % samples only, the rims were accompanied by numerous acicular dendrites growing outward from the overgrowths into the surrounding liquid. In any other matter besides the occurrence of dendrites, the general outlook of the

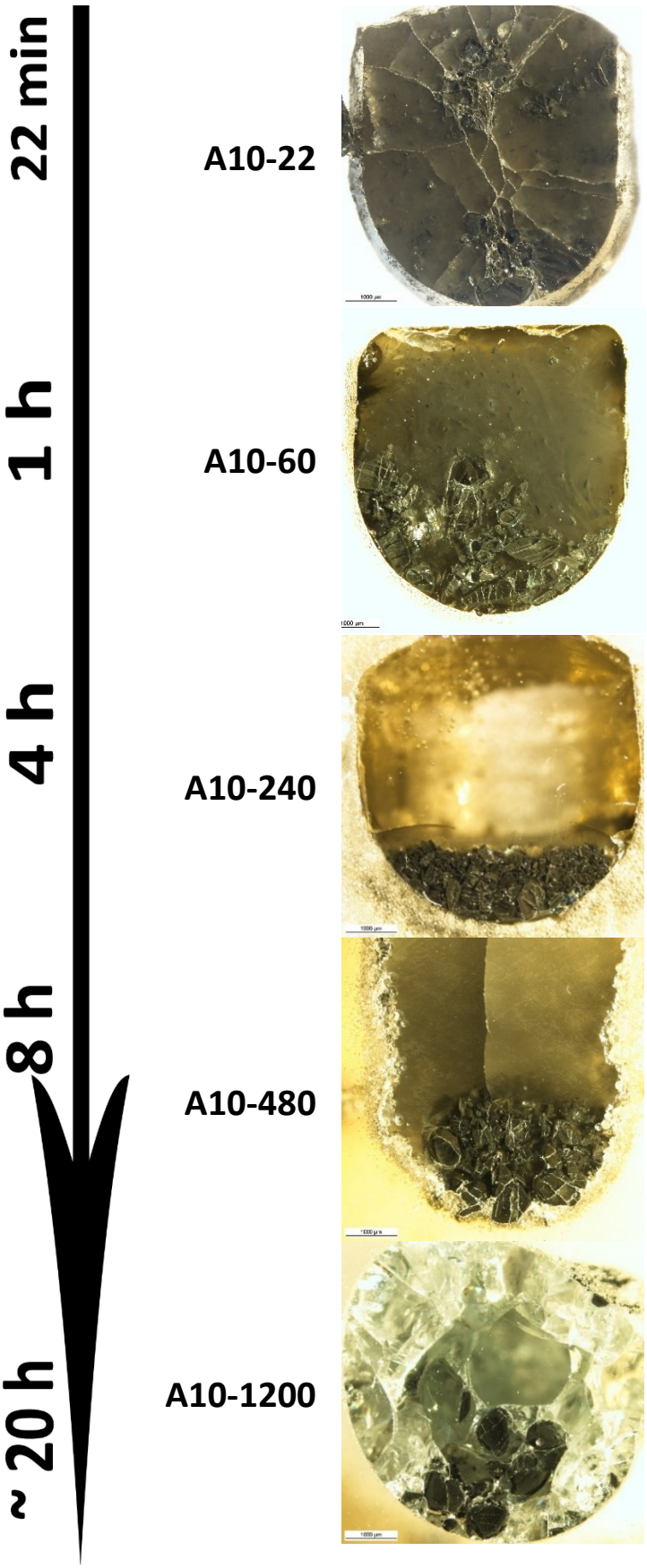


Figure 16 *– A formation of redeposition cumulate from a suspension of olivine crystals, documented by photographs from an optical microscope (2x10 magnification). The scale bar corresponds to 1 mm.*

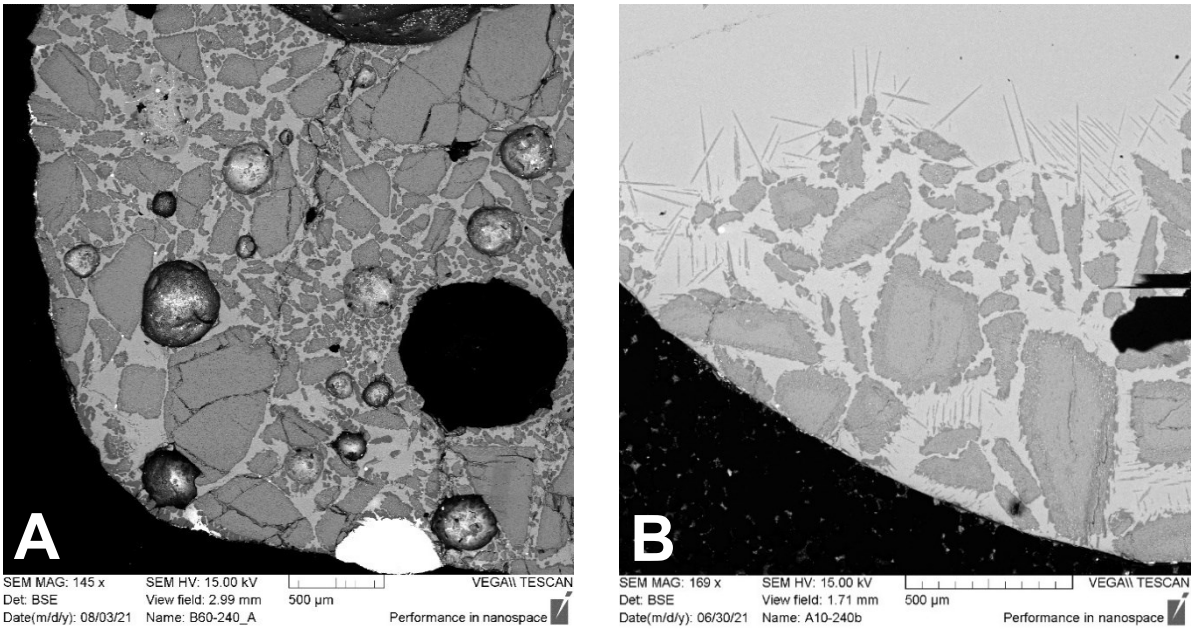


Figure 17 – a) *BSE photograph of a bottom left section of the sample B60-240 (4-hour run), showing a moderately large crystal-filled area with several trapped bubbles.* ***b)*** *BSE photograph of a prominently size-sorted cumulate layer at the bottom of sample A10-240 (4-hour run), with a large number of dendritic crystals at the interface. The sample is also included in Fig. 16.*

rims does not appear to be dependent on the quench method, initial crystallinity, or duration of the experiment in any way. This consistence regardless of the run duration supports an assumption the rims grew during the quench of the sample.

However, the shape of crystals themselves does seem to evolve through time. When the grains from 20 minutes long experiments were compared with those that underwent longer runs, we noted a slight but progressive rounding of the sharp edges (Fig. 19). It is important to note however that the conspicuousness of this effect might be significantly dependent on the cutting plane of the grain, as in some instances a crystal appeared perfectly rounded on one side with a prominently sharp corner on the other. Comparison of the samples have also shown an obvious change in crystal size taking place in 10 wt. % ol. experiments as the time progresses. With prolonging duration of the run, grains with initially larger size visibly grow and the number of small fragments accompanying them is decreasing. In turn, high-crystallinity samples did not seem to be substantially affected by this process as the general layout of the crystal framework was consistent throughout both experimental series.

VI.2.2. Textural data

Resulting values calculated by R software (R Core team 2020) (Tab. 4) were plotted against time to visualize the progressive changes of the texture, namely the development of crystallinity (Fig. 19), spatial distribution of grains defined by the clustering index R (Jerram et al 1996, Fig. 21) and crystal growth measured by median grain size *L* (Fig. 20).

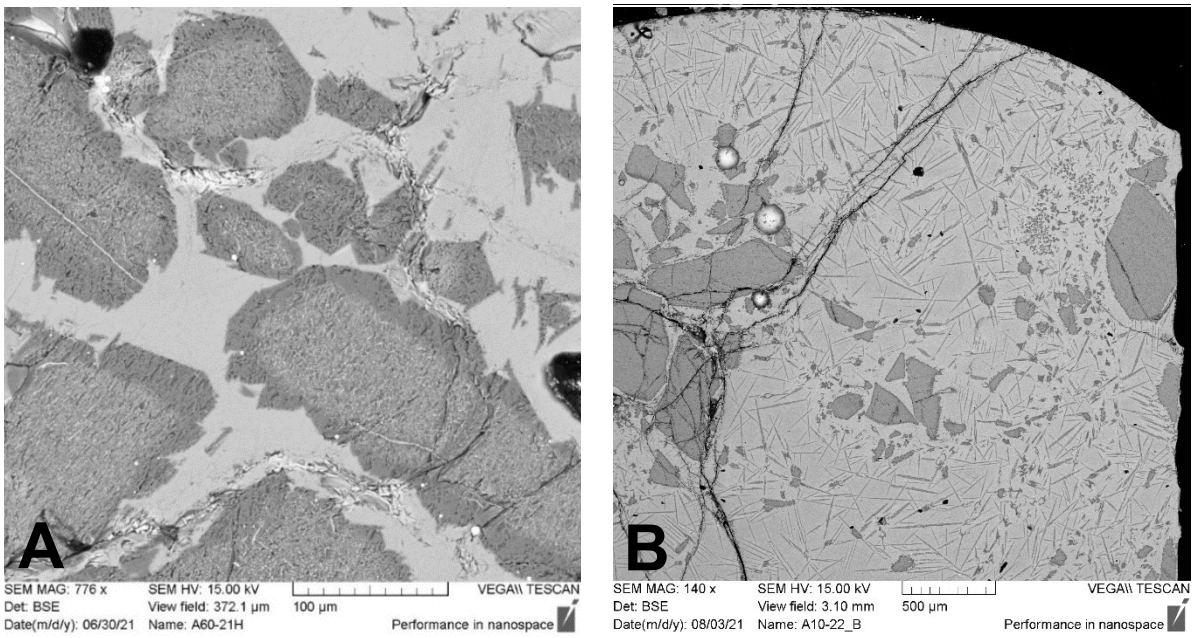


Figure 18 *– **a)** A close-up BSE photograph of olivine crystals from sample A60-1200, showing the rounded shape of grains after 20 hours in the furnace. The light grey centres are the original crystals, the darker irregularly shaped overgrowths are the quench rims.* ***b)*** *A BSE photograph of the upper part of the sample A10-22, showing the uneven distribution of crystals in the capsule after 20 minutes. Grains have shown a tendency to form clusters, mainly in the central part of the sample.*

Table 4 – A complete list of performed experiments along with the textural data.

Sample	*T*	Φ_0 (vol.%)	*t*	*N*	Area (µm²)	Φ (vol.%)	Area density (1/mm²)	L_{MFer} (µm)	*R*
A10-22	1390	9.5	22	253	16404355	13.4481	15.4227	53.2655	0.86460
AA10-22	1390	9.5	22	302	15298435	10.1961	19.7405	41.1211	0.84373
A10-60	1390	9.5	60	144	10167188	32.4351	14.1632	94.7462	1.13520
A10-240	1390	9.5	240	169	3341591	51.4320	50.5747	79.9359	1.14932
A10-480	1390	9.5	480	68	5525393	42.7961	12.3068	127.0861	1.11514
A10-1200	1390	9.5	1200	17	3831166	39.7960	4.4372	271.7444	1.35875
A60-22	1390	58.6	22	224	5191149	55.7497	43.1503	58.1527	1.01045
A60-60	1390	58.6	60	104	2614942	61.6461	39.7714	70.3521	1.07663
A60-240	1390	58.6	240	466	10146042	60.7734	45.9292	58.8322	1.04001
A60-1200	1390	58.6	1200	68	1025591	39.8321	66.3032	62.3359	1.19147
B10-22	1350	9.5	22	381	16356406	7.2834	23.2936	35.8626	1.01522
B10-240	1350	9.5	240	255	9555388	27.8262	26.6865	74.1357	1.25370
B60-22	1350	58.6	22	551	5081791	53.8649	108.4263	41.1884	1.03189
B60-240	1350	58.6	240	404	6076260	50.5060	66.4882	52.3848	1.11587

List of variables*: T – temperature (°C), Φ_0 – initial crystallinity (vol. %), Φ – measured crystallinity, t – time (min), N – a number of grains in the sample, L_{MFer} – median crystal size measured as a maximum Feret's diameter, R – the clustering index. Area density is a number of crystals per unit area in a studied section.*

Crystallinity, Φ

When plotted against time, crystallinity of the A10 samples shows a sharp increase in the first four hours, corresponding to the period when crystals move through the capsule to form a cumulate layer on the floor (Fig. 16). However, following the accumulation stage, crystallinity considerably drops in the next four hours and then proceeds to mildly decrease toward the 20-h timestamp. Crystallinity of the A60 samples undergoes a similar development, where the early increase stops after one hour and is followed by a decline, bringing the crystallinity lower than its initial value after 20 hours. Even though series B covers a significantly smaller time period, the behaviour of crystallinity generally corresponds to the results from series A; For the liquid-rich environment, crystallinity steadily rises over the duration of four hours, while the crystal-filled experiments show a modest decrease.

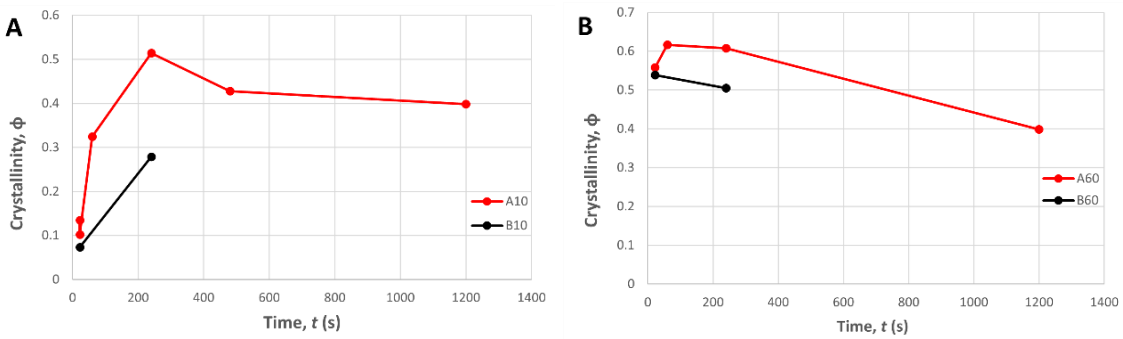


Figure 19 *– Crystallinity of the samples plotted against time, with series A marked as red lines and series B as black.* ***a)*** *10 wt. % ol. samples,* ***b)*** *60 wt. % ol. samples.*

Crystal size, L

Similarly to the previous case, the evolution of median crystal size greatly differs between the crystal-rich and liquid-rich samples. In case of the A10 samples, the crystal size sharply rises from the initial median value of 2.9 um to approximately 50 µm in under one hour, and then continues to linearly increase up to 270 µm after 20 hours. The B10 samples behave alike, and their measured values of median crystal sizes are roughly the same. In the high-crystallinity samples, the size also increases rapidly up to nearly 60 µm after a short time period and continues to rise up to 70 µm in one hour, however, a following development shows a drop back to the 60 µm and a continuous stagnation on this number. The sudden spike between three identical values caused by sample A60-60 might be attributed to the unrepresentative nature of the sample due to severe breakage and absence of larger crystal-filled areas. Overall, no significant increase in crystal size took place in the 20-hour time period following the initial rise.

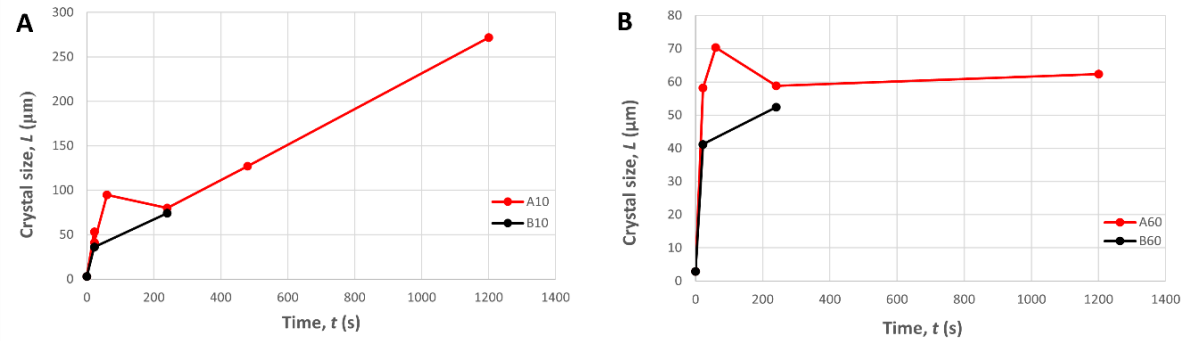


Figure 20 *– The evolution of median crystal size with time. Series A marked as red lines, series B as black.* ***a)*** *10 wt. % ol. samples,* ***b)*** *60 wt. % ol. samples.*

Clustering index, R

The clustering index *R*, defining a spatial relationship between a population of crystals, has been calculated by measuring the distance between a centre if each grain and the centre of its nearest neighbour, followed by dividing the distances by their expected value in a system with a random distribution of grains (Jerram et al 1996). For the A10 samples, the first value of ~0.85 given by two 20-min long runs has shown that crystals are prone to form clusters, which confirmed our observation from the back scattered electron images. The three following samples representing the textural development in the next 8 hours have all provided values of approximately 1.1, meaning that the crystals tend to occur in an ordered pattern instead of continuing to cluster. The *R* value then continues to rise up to 1.3 for the longest run, however, given by the small number of crystals in this sample (*N*=17), this result bears a significant statistical error (Jerram et al 1996).

A60 samples show a roughly similar behaviour to their low-crystallinity counterparts. Starting with a random distribution of grains after 20 minutes (*R*≈1.01), the value of *R* then rises and stagnates around 1.06 and subsequently increases further up to ~1.2 after 20 hours. Finally, neither B10 of B60 samples deviate from this behaviour pattern.

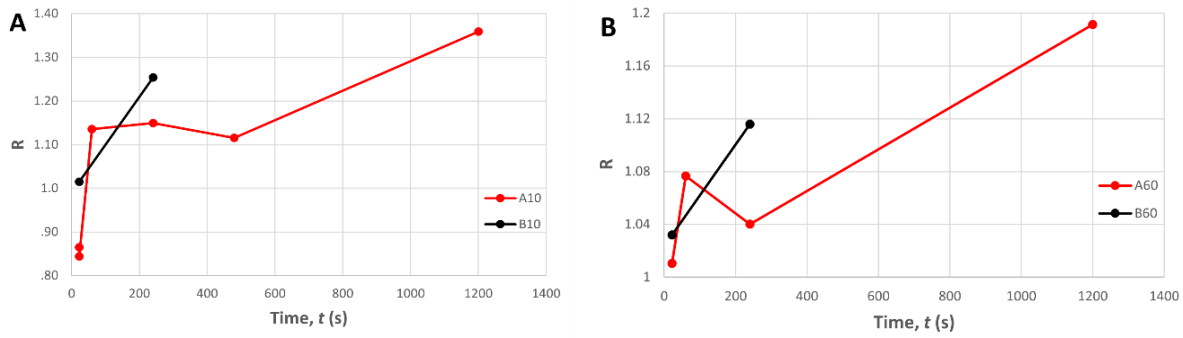


Figure 21 *- Evolution of clustering index (R) values in our samples over a 20h period. Series A marked as red lines, series B as black.* ***a)*** *10 wt. % ol. samples,* ***b)*** *60 wt. % ol. samples.*

VI.3 Discussion

VI.3.1. Mechanical behaviour of crystal-liquid suspensions

All but one of our experiments show a rise of crystallinity in the first four hours, with the sharpest increase observed in the liquid-rich sets of experiments A10 and B10. For these samples, the four-hour time period corresponds to the stage of progressive crystal accumulation at the floor of the capsule, meaning that the rise in crystallinity most likely results from the crystal settling. At its highest point, the crystallinity of the cumulate pile reaches 50 vol. %, a slightly higher value than what has been measured in settling experiments by Schmidt et al (2012). As the movement of crystals in 60 wt. % ol. samples is restricted by their high initial crystallinity, its further rise is considerably less steep than in the case of their 10 wt. % counterparts. Although it is quite possible that the crystal framework did slightly settle in the incipient phases of the experiment, we assume that the crystallinity rise in A60 samples is rather caused by expansion of bubbles inside the capsule, forcibly redistributing the material towards its sides and pressing the crystals closer.

In both instances, the increase of crystallinity has stopped at similar values of ~50 and ~60 vol. percent respectively, which aligns well with a commonly used value of 55 % as the rigid percolation threshold (RPT) (e.g., Marsh 1981, 1989). Above this content of solid particles, crystals form a closely packed solid framework and resist further movement, which is why the crystallinity of our accumulated crystal piles did not continue to mechanically increase beyond this limit. In the A60 samples, the increase of crystallinity above the limit reached by the 10 wt. % experiments (as well as above the RPT) can again be explained by the pressure of expanding bubbles, inducing an effective filter-pressing of the crystal framework (Sisson and Bacon 1999).

A peculiar reversal of development occurs after the initial crystallinity increase, where it begins to generally decrease in both the A10 and A60 samples towards approximately 40 vol. % throughout the next 16 hours. This observation might be a result of statistic fluctuations due to poor handling of the starting mixture or an analysis of non-representative area of the sample. However, we are inclined to believe that the drop of crystallinity is caused by a dissolution of olivine given by phase relations in the system. Although the employed olivine crystals were of natural composition and therefore contained a certain amount of iron, we did not consider its effect while choosing the experimental temperature and treated the mixture as pure Di-Fo-An system. Consequently, the iron content likely lowered the liquidus temperature and crystals began to dissolve. Interestingly enough, the decrease of crystallinity and therefore the effect of dissolution becomes noticeable only after the four hours of continuous increase, even though it is highly improbable that the dissolution did not occur in the system since the beginning. This shows that the effect of settling (A10 samples) and filter-pressing (A60 samples) overran the dissolution which thus, being the much slower process, become dominant only once the mechanical rearrangements ended.

Furthermore, we suspect that dissolution have caused the series B samples to behave differently from what was expected. The purpose of series B has been to invoke adcumulus growth of crystals by lowering the temperature by 40 °C, however, as dissolution likely operated in the system even under these conditions, any adcumulus overgrowths were lost to its effect. Dissolution have also likely acted as a contributing factor to the progressive changes in crystal shape, observed on the BSE photographs (Fig. 19). Although the effects of equilibration on the sharp crystal edges would be the same, dissolution might be the prevailing process in our case, given the short time frame and the extend of changes (e.g., Cabane et al 2005; see below).

VI.3.2. Textural evolution of cumulates

All series of experiments show a rapid increase of median crystals size by an order of magnitude in under one hour. When compared to other experimental works, such as the study of Ostwald ripening by Cabane et al (2005), the growth observed in our samples is considerably quicker; Cabane et al (2005) measured an increase of crystal size from 5 µm to 50 µm after ~500 hours, while we achieved a rise in median size from 2.9 µm to 90 µm in one hour (A10 series). Similarly, in a study of olivine coarsening in a haplobasaltic system by Park and Hanson (1999), an increase from 7 µm to 28 µm was gained in approximately 10 days. This staggering difference in the rate of crystal growth has led us to assume that the driving force of size increase in our experiments is likely not the Ostwald ripening, instead it results from the effect of simple dissolution of smaller grains while the larger ones persist.

Although both the A10 and A60 experiments exhibit a steep initial rise in crystal size, the two types of suspension begin to show a contrasting behaviour as the time progresses. While the median crystal size in 10 wt. % experiments increases linearly towards 300 µm under 20 hours, the grains in high-crystallinity samples do not continue to grow after reaching ~70 µm in four hours. We suggest that this dichotomy can be explained as an influence of compositional convection (Tait et al 1994). As the crystals in liquid-rich experiments can not only move through its capsule, but also have access to the overlaying melt once they have settled to the floor, the interstitial liquid inside the cumulate might be regularly replaced by compositional convection from the bulk melt reservoir above, keeping the conditions around crystals identical to when they moved freely and enabling them to grow linearly. In contrast, the locked crystal framework in high-crystallinity samples causes the growth to stagnate after approximately one hour, as there is no convection to supply the grains with fresh melt.

The clustering index generally rises steadily in all performed experiments, with a more significant increase in 10 wt. % samples. In both suspension types, the initial R value of ~1 implies a random distribution of grains, which then proceeds to evolve towards an anti-clustered (i.e., ordered) texture with the maximum values of 1.19 (A60) and 1.3 (A10) after 20 hours. The slower rise of R in A60 samples is in accordance with its general tendencies to increase with time during crystal accumulation (e.g., Jerram et al 2003, Špillar and Dolejš 2015), consequently favouring our low-crystallinity

suspensions in terms of the ordering rate. The initial clustering observed in A10-22 samples (Fig. 18b) might have been caused by heterogeneity of the starting olivine-glass mixture, but as it does not affect the further development of the texture, we do not attribute much significance to it.

VI. Conclusions

Cumulate rocks are formed by a sequence of processes involving movement of crystals and their surrounding liquid, whose common denominator is a fractional crystallization of the parental magma (Irvine 1982). Given the interconnectedness of numerous mechanisms operating inside mafic magma chambers, cumulates present a widely variable group of rocks in terms of texture, modal composition, and chemistry (Sparks et al 1985, Hunter 1996). In this work, we attempted to experimentally demonstrate some of the processes considered as crucial during cumulate formation, and texturally evaluate the resulting samples as means to quantify the observed changes. The performed set of high-temperature experiments in a haplobasaltic system, tracking a development of two types of olivine-liquid suspensions over a period of 20 hours, has shown the following:

- Crystallinity of the samples rises steadily with progressive accumulation of crystals up to a maximum value of ~50 %, which corresponds well to the general estimates of rigid percolation threshold (Marsh 1981, 1989) as well as to experimental studies of initial porosity of cumulates formed by crystal redistribution (e.g., Wager et al 1960). However, the crystallinity begins to drop as the mechanical movement of crystals stops, which seems to be caused by dissolution of olivine in our experiments due to their Fe-content. As the accumulation is a considerably faster process, the effect of dissolution appears as delayed.
- An increase of median grains size with time has been considerably faster in our set of experiments than in comparable studies of coarsening in basaltic melt-olivine systems (Park and Hanson 1999, Cabane et al 2005). We attribute this disparity to a simple dissolution of smaller grains in the system rather than to coarsening of the larger ones, again due to lower liquidus temperature of the olivine grains.
- In liquid-rich environments, the rate of crystal growth is not affected by their progressive accumulation on the floor of the reservoir and the median grain size continues to enlarge linearly long after the cumulate layer has formed. Meanwhile, in a closely packed crystal framework, the growth begins to stagnate after a short initial period of increase. This contrasting behaviour can be explained as an effect of compositional convection (Tait et al 1994) which drives an exchange of liquid between the cumulate and the bulk melt but proves to be ineffective in high-crystallinity conditions.
- The clustering index *R* (Jerram et al 1996) increases with time in all our experiments pointing to progressive ordering (anti-clustering) of olivine suspension. While this effect if expectable in liquid-rich and settling-dominated environment (e.g., Špillar and Dolejš 2015), it is surprising in high-crystallinity conditions.

- An accessory series of experiments performed under lower temperatures has shown no signs of adcumulus growth in either environment, instead the samples display an identical behaviour as the main experimental series. The absence of adcumulus overgrowths is likely a result of dissolution taking place even under the lower temperature conditions.

References

Atkins, J.E. & McBride, E.F. (1992). Porosity and packing of Holocene River, dune and beach sands. American Association of Petroleum Geology, 76. 339-355.

Barnes, S.J. & Mole, D.R. & Le Vaillant, M. & Campbell, M.J. & Verrall, M.R. & Roberts, M.P. & Evans, N.J. (2016). Poikilitic textures, heteradcumulates and zoned orthopyroxenes in the Ntaka Ultramafic Complex, Tanzania: Implications for crystallization mechanisms of oikocrysts. Journal of Petrology, 57. 1171-1198.

Bédard, J.H. (2015). Ophiolitic magma chamber processes, a perspective from the Canadian Appalachians. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp. 693-732.

Bottinga, Y. & Weill, D.F. (1970). Densities of liquid silicate systems calculated from partial molar volumes of oxide components. American Journal of Science, 269. 169-182.

Boudreau, Alan & McBirney, Alexander (1997). The Skaergaard Layered Series. Part III. Non-dynamic Layering, Journal of Petrology, 38. 1003-1020.

Boudreau, Alan (2011). The evolution of texture and layering in layered intrusions. International Geology Review, 53. 330-353.

Brandeis, G & Jaupart, C. & Allègre, C. J. (1984). Nucleation, crystal growth and the thermal regime of cooling magmas, Journal of Geophysical Research, 89. 10161-10177.

Bryan, Scott & Ernst, Richard (2008). Revised definition of Large Igneous Provinces (LIPs). Earth-Science Reviews, 86. 175-202.

Cabane, H. & Laporte, D. & Provost, A. (2001). Experimental investigation of the kinetics of Ostwald ripening of quartz in silicic melts. Contributions to Mineralogy and Petrology, 142. 361-373.

Cabane, H. & Laporte, D. & Provost, A. (2005). An experimental study of Ostwald ripening of olivine and plagioclase in silicate melts: Implications for the growth and size of crystals in magmas. Contributions to Mineralogy and Petrology, 150. 37-53.

Campbell, I.H. (1978). Some problems with the cumulus theory. Lithos, 11. 311-323.

Campbell, I.H. (1982). Layered intrusions: a mini review. Workshop on Magmatic Processes of Early Planetary Crusts: Magma Oceans and Stratiform Layered Intrusions. 62-65. LPI Tech Rept Houston.

Campbell, I.H. (1987). Distribution of orthocumulate textures in the Jimberlana Intrusion. Journal of Geology, 95. 35-54.

Cawthron, R.G. & Ashwal L.D. (2009). Origin of anorthosite and magnetitite layers in the Bushveld Complex, constrained by major element compositions of plagioklase. Journal of Petrology, 50. 1607-1637.

Cawthorn, R.G. (2015). The Bushveld Complex, South Africa. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 517-588.

Chistyakova, Sofia & Latypov, Rais (2014). Adcumulate mafic dykes in layered intrusions: A case study of a late-stage dyke in the Bayantsagaan layered intrusion, Mongolia. Geological Magazine, 152. 621-631.

Cooper, R.F. & Kohlstedt, D.L. (1986). Rheology and structure of olivine basalt partial melts. Journal of Geophysical Research, 91. 9315-9323.

Eales, H.V. (2000). Implications of the chromium budget of the Western Limb of the Bushveld Complex. South African Journal of Geology, 103. 141-150.

Ernst, Richard & Jowitt, Simon & Blanchard, J & Liikane, D. (2017). Large Igneous Provinces and Their Mafic-Ultramafic Intrusions. IOP Conference Series: Earth and Environmental Science, 110.

Fediuk, F. (1971). Ultramafity krkonošsko-jizerské oblasti. Acta Universitatis Carolinae, Geologica, 4. 319-343.

Gray, N.H. & Philpotts, A.R. & Dickson, L.D. (2003). Quantitative measures of textural anisotropy resulting from magmatic compaction illustrated by a sample from the Palisades sill, New Jersey. Journal of Volcanology and Geothermal Research, 121. 293-312.

Hess, H.H. (1960). Stillwater igneous complex. Geological Society of America Memoirs, 80. 1-230.

Hersum, T. & Hilpert, M. & Marsh, B. (2005). Permeability and melt flow in simulated and natural partially molten basaltic magmas. Earth and Planetary Science Letters, 237. 798-814

Higgins, M.D. (2002). A crystal size-distribution study of the Kiglapait layered mafic intrusion, Labrador, Canada: evidence for textural coarsening. Contributions to Mineralogy and Petrology, 144. 314-330.

Higgins, M.D. (2011). Textural coarsening in igneous rocks. International Geology Review, 53. 354-376.

Higgins M.D. (2015). Quantitative Textural Analysis of Rocks in Layered Mafic Intrusions. In: Charlier B., Namur O., Latypov R., Tegner C. (ed.) Layered Intrusions. Springer Geology. Springer, Dordrecht, pp 153-181.

Holness, M.B. & Hallworth, M. A. & Woods, A. & Sides, R. E. (2007). Infiltration metasomatism of cumulates by intrusive magma replenishment: the Wavy Horizon, Isle of Rum, Scotland. Journal of Petrology, 48. 563-587.

Holness, M.B. & Tegner, C. & Nielsen, T.F. (2009). Constraining the thickness of the crystal mush in layered mafic intrusions. American Geophysical Union, Fall Meeting 2009, abstract V13F-05.

Holness, M.B. Vernon R.H. (2015). The influence of interfacial energies on igneous microstructures. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 183-228.

Holness, M.B. & Nielsen, Troels & Tegner, Christian (2017). The Skaergaard Intrusion of East Greenland: Paradigms, Problems and New Perspectives. Elements, 13. 391-396.

Holness, M.B. & Vukmanovic, Z. & Mariani, E. (2017). Assessing the role of compaction in the formation of adcumulates; a microstructural perspective. Journal of Petrology, 58. 643-674.

Holness, M.B. & Clemens, J.D. & Vernon, R.H. (2018). How deceptive are microstructures in granitic rocks? Answers from integrated physical theory, phase equilibrium, and direct observations. Contributions to Mineralogy and Petrology, 173. 62.

Holness M.B. & Morris C. & Vukmanovic Z. & Morgan D.J. (2020). Insights into magma chamber processes from the relationship between fabric and grain shape in troctolitic cumulates. Frontiers in Earth Science, 8. 352-370.

Holness, M. B. (2020). Cumulates and Layered Igneous Rocks. In: Alderton D., Elias A.S. Encyclopedia of Geology (2nd ed.). Academic Press, pp 99-112.

Humphreys, M.C.S. (2009). Chemical evolution of intercumulus liquid, as recorded in plagioklase overgrowth rims from the Skaergaard Intrusion. Journal of Petrology, 50. 127-145.

Hunter, R.H. (1987). Textural equilibrium in layered igneous rocks, in Parsons, I., (ed.) Origins of igneous layering: Boston, D. Reidel, pp 473-503.

Hunter, R.H. (1996). Textural development in cumulate rocks. In: Cawthorn, G.A. (ed.) Layered Intrusions. Elsevier, Amsterdam, pp 77-101.

Huppert, H. & Sparks, R.S. (1980) The fluid dynamics of a basaltic magma chamber replenished by influx of hot, dense ultrabasic magma. Contributions to Mineralogy and Petrology, 75. 279-289.

Irvine, T.N. & Smith, C.H. (1967). The ultramafic rocks of the Muskox intrusion, Northwest Territories, Canada. In: Wyllie PJ (ed.) Ultramafic and related rocks. Wiley, New York, pp 38-49.

Irvine, T.N. (1974). Petrology of the Duke Island, ultramafic complex in southeastern Alaska. Geological Society of America Memoirs, 138. 1-240.

Irvine, T.N. (1980). Magmatic infiltration metasomatism, double-diffusive fractional crystallisation and adcumulus growth in the Muskox intrusion and other layered intrusions, in: Hargraves, R.B., (ed.), Physics of magmatic processes. Princeton University Press, pp 325-384.

Irvine, T.N. (1980a). Magmatic density currents and cumulus processes. Americal Journal of Science, 280. 1-58.

Irvine, T.N. (1982). Terminology for Layered Intrusions. Journal of Petrology, 23. 127-162.

Irvine, T.N. (1987) Appendices. Parsons I. (eds) Origins of Igneous Layering. NATO ASI Series (Series C: Mathematical and Physical Sciences), 196. 641-656.
Irvine, T.N. & Andersen J.C.O. & Brooks C.K. (1998). Included blocks (and blocks within blocks) in the Skaergaard intrusion: geologic relations and the origins of rhythmic modally graded layers.
Geological Society of America Bulletin, 110. 1398-1447.
Jackson, E.D. (1970). The cyclic unit in layered intrusions - a comparison of the repetitive stratigraphy in the ultramafic parts of the Stillwater, Muskox, Great Dyke and Bushveld Complexes. Spec. Publ. geol. Soc. S. AJr, 1. 391-423.
Jackson, E.D. (1967). Ultramafic cumulates in the Stillwater, Great Dyke, and Bushveld intrusions. In: Wyllie PJ (ed.) Ultramafic and related rocks. Wiley, New York, pp 20-38.
Jenkins, M. & Mungall, James (2018). Genesis of the Peridotite Zone, Stillwater Complex, Montana, USA. Journal of Petrology, 59. 2157-2189.
Jerram, D. & Cheadle, M. & Hunter, R. & Elliott, M. (1996). The spatial distribution of grains and crystals in rocks. Contributions to Mineralogy and Petrology, 125. 60-74.
Keiding, Jakob & Tegner, Christian & Brooks, Kent & Kent, Adam & Lesher, Charles & Nielsen, Troels & Wiedenbeck, Michael (2010). Parental magma of the Skaergaard intrusion: Constraints from melt inclusions in primitive troctolite blocks and FG-1 dykes. Contributions to Mineralogy and Petrology, 159. 61-79.
Latypov, R.M. & Chistyakova, S.Y. & Mukherjee, Ria (2017). A novel hypothesis for origin of massive chromitites in the Bushveld igneous Complex. Journal of Petrology, 58. 1899-1940.
Latypov, R.M. & Chistyakova, S.Y. (2020) Origin of non-cotectic cumulates: A novel approach. Geology, 48. 604-608.
Lee C. A. (1996). A Review of Mineralization in the Bushveld Complex and some other Layered Intrusions. Developments in Petrology, 15. 103-145.
Maier, Wolfgang (2005). Platinum-group element (PGE) deposits and occurrences: Mineralization styles, genetic concepts, and exploration criteria. Journal of African Earth Sciences, 41. 165-191.
Maier, Wolfgang & Barnes, Sarah-Jane (2008). Platinum-group elements in the UG1 and UG2 chromitites, and the Bastard reef, at Impala platinum mine, western Bushveld Complex, South Africa: Evidence for late magmatic cumulate instability and reef constitution. South African Journal of Geology, 111. 159-176.
Marks, Michael A. W. & Markl, Gregor (2015). The Ilímaussaq Alkaline Complex, South Greenland. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 649-691.
Marsh, B.D. (1981). On the crystallinity, probability of occurrence, and rheology of lava and magma. Contributions to Mineralogy and Petrology, 78. 85-98.
Marsh, B.D. (1989). Magma chambers. Annual Review of Earth and Planetary Sciences, 17. 439-474.
McBirney, Alexander & Noyes, R.M. (1979). Crystallization and layering of the Skaergaard intrusion. Journal of Petrology, 20. 487-553.
McBirney, A.R. & Sonnenthal, E.L. (1990). Metasomatic replacement in the Skærgaard Intrusion, East Greenland: Preliminary observations. Chemical Geology, 88. 245-260.
McBirney, A.R. & Hunter, R.H. (1995). The cumulate paradigm reconsidered. Journal of Geology, 103. 114-22.
McBirney, A.R. (1996). The Skaergaard Intrusion. In: Cawthorn, R.G. (ed.) Layered Intrusions. Elsevier, Amsterdam, pp 147-180.
McBirney, Alexander & Nicolas, Adolphe (1997). The Skaergaard Layered Series. Part II. Magmatic flow and Dynamic Layering. Journal of Petrology, 38. 569-580.
McCallum, I.S. (1996). The Stillwater complex. In: Cawthorn R.G. (ed.) Layered Intrusions. Elsevier, Amsterdam, pp 441-484
McKenzie, D. (1984). The generation and compaction of partially molten rock. Journal of Petrology, 25. 713-765.
McKenzie, D. (2011). Compaction and crystallization in magma chambers: towards a model of the Skaergaard Intrusion. Journal of Petrology, 52. 905-930.

Meurer, W. & Boudreau, A. (1998). Compaction of igneous cumulates part II: compaction and the development of igneous foliations. Journal of Geology, 106. 293-304.
Morse, Stearns A. (1979b). Kiglapait Geochemistry II: Petrography, Journal of Petrology, 20. 591-624.
Morse, Stearns A. (1986). Convection in aid of adcumulus growth. Journal of Petrology, 27. 1183-1214.
Morse, Stearns A. (2015). Kiglapait Intrusion, Labrador. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 589-648.
Namur, Oliver & Charlier, B. & Pirard, C. & Hermann, J. & Liégeois J.P. & Auwera, J.V. (2011a). Anorthosite formation by plagioclase flotation in ferrobasalt and implications for the lunar crust. Geochimica et Cosmochimica Acta, 75. 4998-5018.
Namur, Olivier & Abily, Bénédicte & Boudreau, Alan & Blanchette, Francois & Bush, John & Ceuleneer, Georges & Charlier, Bernard & Donaldson, Colin & Duchesne, J.C. & Higgins, M. & Morata, D. & Nielsen, Troels & O'Driscoll, B. & Pang, Kwan-Nang & Peacock, Thomas & Spandler, Carl & Toramaru, Atsushi & Veksler, Ilya (2015). Igneous Layering in Basaltic Magma Chambers. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 75-152.
Naslund, H.R. (1984). Supersaturation and crystal growth in the roof-zone of the Skaergaard magma chamber. Contributions to Mineralogy and Petrology, 86. 89-93.
Naslund, H.R. & McBirney, A.R. (1996). Mechanisms of formation of igneous layering. In: Cawthron RG (ed.) Layered Intrusions. Elsevier, Amsterdam, pp 1-43.
Nichols, Gary (2009). Sedimentology and stratigraphy (2nd ed.). Hoboken, NJ: Wiley-Blackwell, p 432.
Paull, C. & Talling, P. & Maier, K. & Parsons, D. & Xu, J. & Caress, D. & Gwiazda, R. & Lundsten, E. & Anderson, K. & Barry, J. & Chaffey, M. & O'Reilly, T. & Rosenberger, K. & Gales, J. & Kieft, Brian & McGann, M. & McCann, M. & Sumner, E. & Clare, M. & Simmons, S. (2018). Powerful turbidity currents driven by dense basal layers. Nature Communications, 9.
Park, Y. & Hanson, B. (1999). Experimental investigation of Ostwald ripening rates of forsterite in the haplobasaltic system. Journal of Volcanology and Geothermal Research, 90. 103-113.
R Core Team (2020). R: A language and environment for statistical computing. R Foundation for Statistical Computing, Vienna, Austria. https://www.R-project.org/.
Sahu, Puspanjali & Jena, M. & Mandre, Narayan R & Venugopal, R. (2020). Platinum Group Elements Mineralogy, Beneficiation, and Extraction Practices – An Overview. Mineral Processing and Extractive Metallurgy Review.
Scoates, James & Wall, Corey (2015). Geochronology of Layered Intrusions. In: Charlier B.L.A., Namur O., Latypov R.M., Tegner C. (ed.) Layered intrusions. Springer, Dordrecht, pp 3-74.
Schmidt, Max & Forien, Mélanie & Solferino, Giulio & Bagdassarov, Nickolai. (2012). Settling and compaction of olivine in basaltic magmas: An experimental study on the time scales of cumulate formation. Contributions to Mineralogy and Petrology, 164.
Sisson, T.W. & Bacon C.R. (1999) Gas-driven filter pressing in magmas. *Geology*, 27. 613-616.
Sparks, R.S.J. & Kerr, R.C. & McKenzie, D.P. & Tait, S.R. (1985). Postcumulus processes in layered intrusions. Geological Magazine, 122. 555-568.
Špillar, Václav & Dolejš, David (2014). Kinetic model of nucleation and growth in silicate melts: Implications for igneous textures and their quantitative description. Geochimica et Cosmochimica Acta, 131. 164-183.
Špillar, Václav & Dolejš, David (2015). Melt extraction from crystal mushes: Numerical model of texture evolution and calibration of crystallinity-ordering relationships. Lithos, 239. 19-32.
Tait, S.R. & Huppert, H.E. & Sparks, R.S.J. (1984). The role of compositional convection in the formation of adcumulus rocks. Lithos, 17. 139-146.
Tait, S. & Jaupart, C. (1992). Compositional convection in a reactive crystalline mush and melt differentiation. Journal of Geophysical Research: Solid Earth, 97. 6735-6756.
Tegner, C. & Thy, P. & Holness, M.B. & Jakobsen, J.K. & Lesher, C.E. (2009). Differentiation and compaction in the Skaergaard Intrusion. Journal of Petrology, 50. 813-840.

Upton, B. G. J. & Parsons, I. & Emeleus, C. H. & Hodson, M. E. (1996). Layered Alkaline Igneous Rocks of the Gardar Province, South Greenland. In: Cawthorn, R.G., (ed.), Layered Intrusions. Elsevier, Amsterdam, pp 331-363.

Verhoeven, J. & Schmalzl, J. (2009). A numerical method for investigating crystal settling in convecting magma chambers. Geochemistry Geophysics Geosystems, 10.

Vukmanovic, Z. & Holness, M.B. & Monks, K. & Andersen, J.C.Ø. (2018). The Skaergaard trough layering: sedimentation in a convecting magma chamber. Contributions to Mineralogy and Petrology, 173.

Wadsworth, W.J. (1961). The layered ultrabasic rocks of south-west Rhum, Inner Hebrides. Philosophical Transactions of the Royal Society of London, 244. 21-64.

Walker, D. & Jurewicz, S. & Watson, E.B. (1985). Experimental observation of an isothermal transition from orthocumulus to adcumulus texture (abs.) EOS, 66. 362.

Wager, L. R. & Deer, W. A. (1939). Geological investigations in East Greenland, Pt. III. The Petrology of the Skaergaard Intrusion, Kangerdlugssuag, East Greenland, Meddelelser om Grønland, 105. 1-352.

Wager, L.R. & Brown, G.M. & Wadsworth, W.J. (1960). Types of igneous cumulates. Journal of Petrology, 1. 73-85.

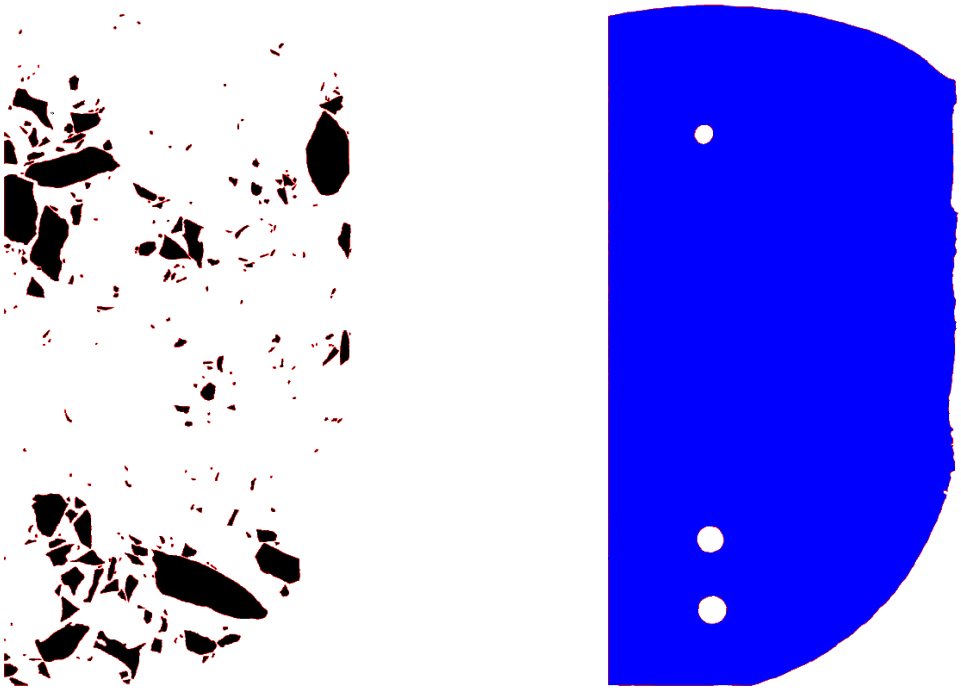
Wager, L.R. (1961). A note on the origin of ophitic texture in the chilled olivine gabbro of the Skaergaard intrusion. Geological Magazine, 98. 353-366.

Wager, L.R. (1963). The mechanism of adcumulus growth in the layered series of the Skaergaard intrusion. Mineralogical society of America, Special Paper, 1. 1-9.

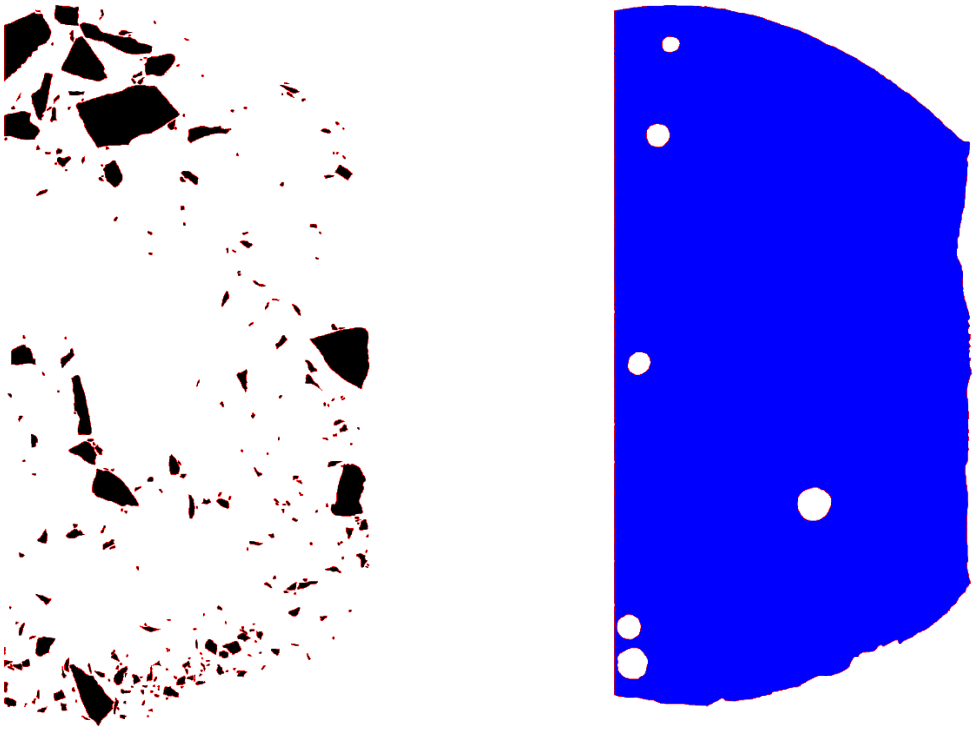
Wager, L.R. Brown, G.M. (1968). Layered Igneous Rocks. Oliver & Boyd, Edinburgh, London. pp 588

Appendix

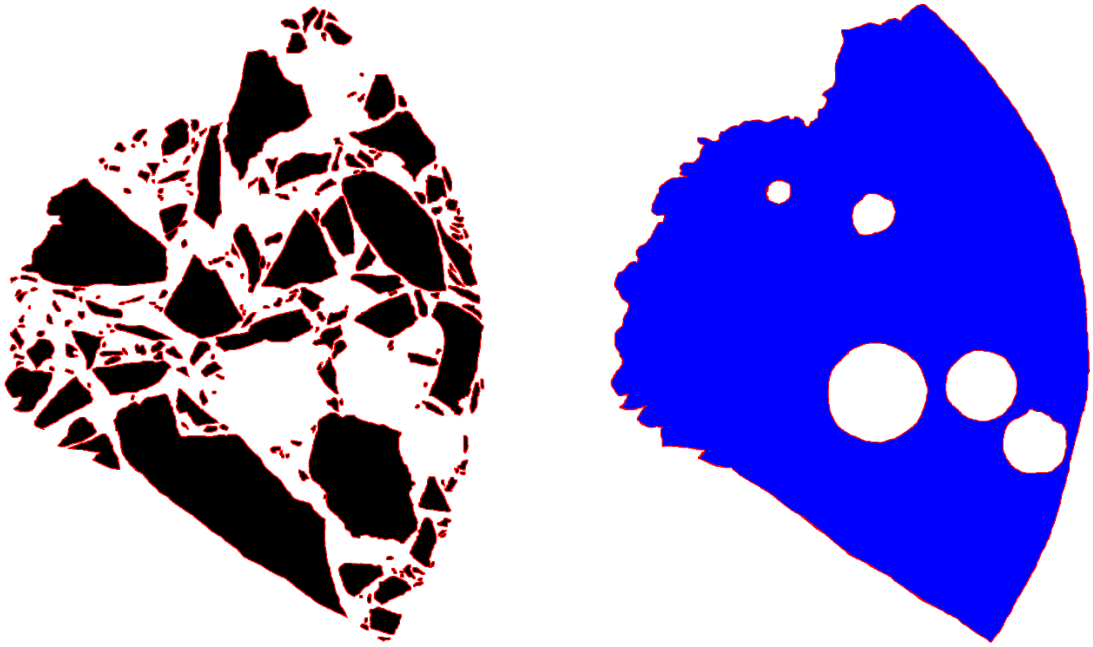
Digitalized textures of studied samples and outlines of studied areas



A10-22

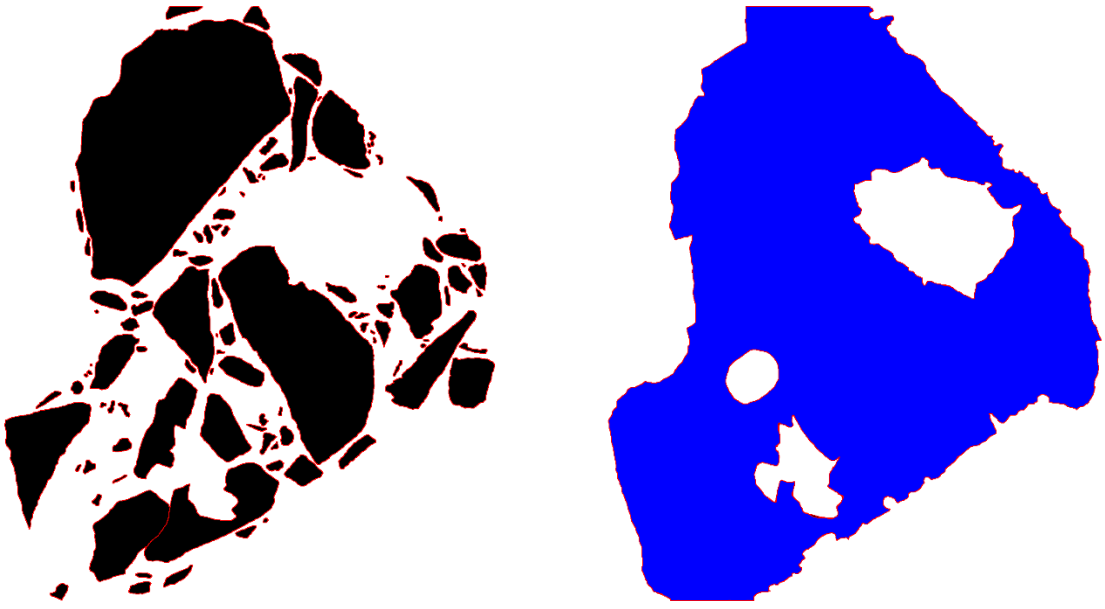


AA10-22

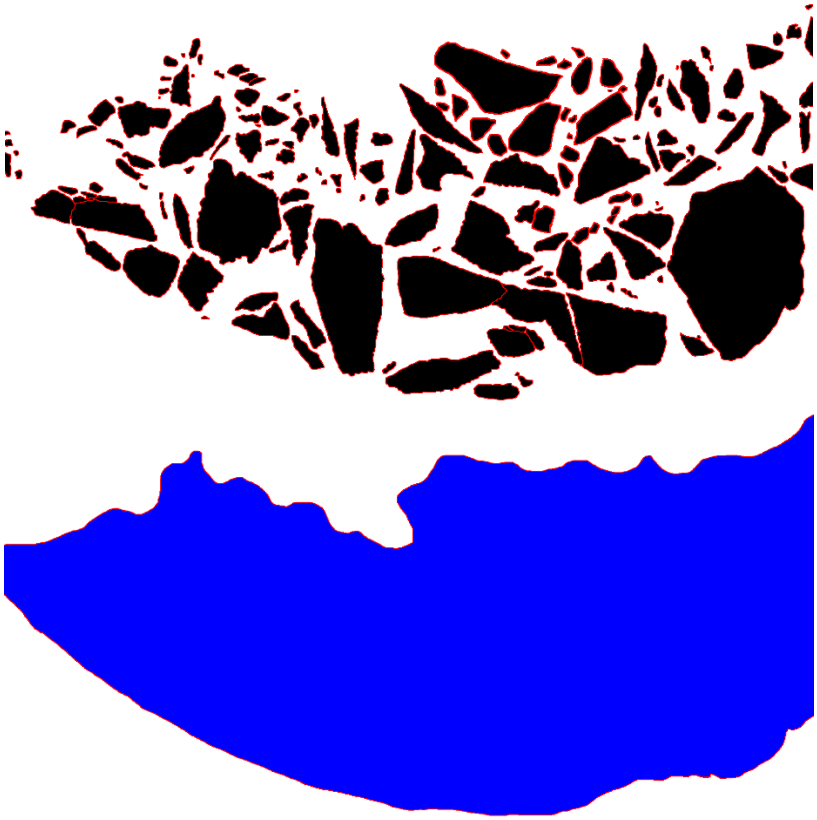


A60-22

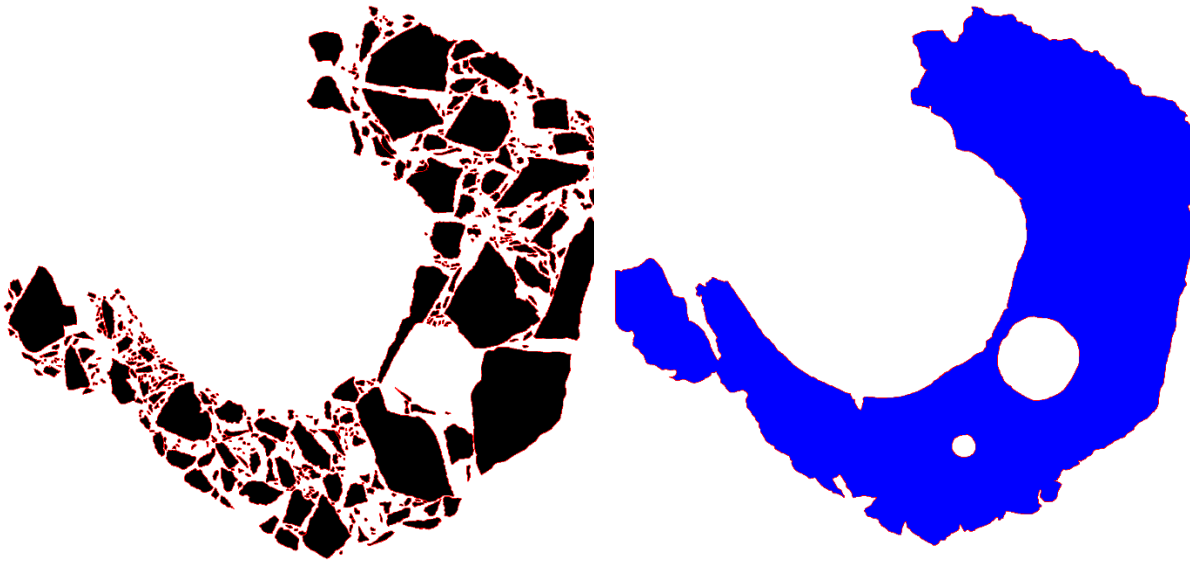
A10-60



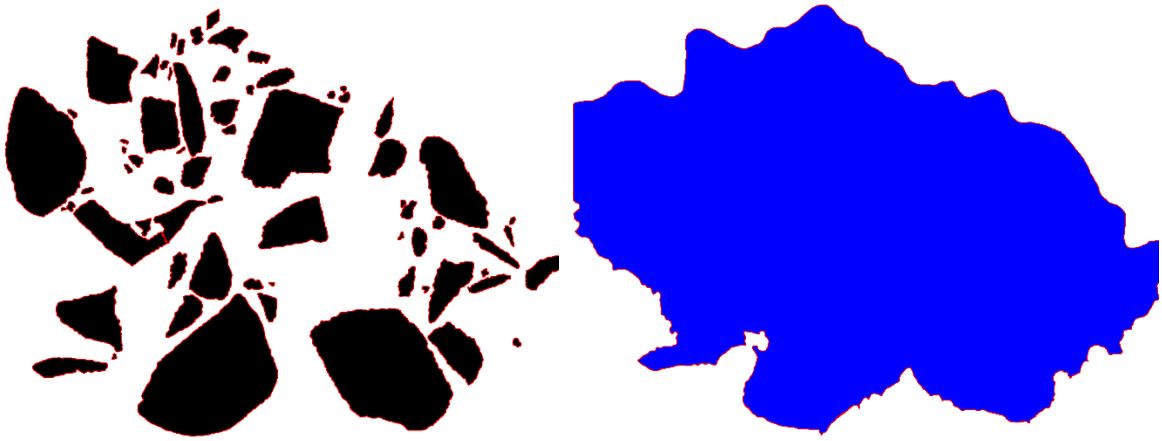
A60-60



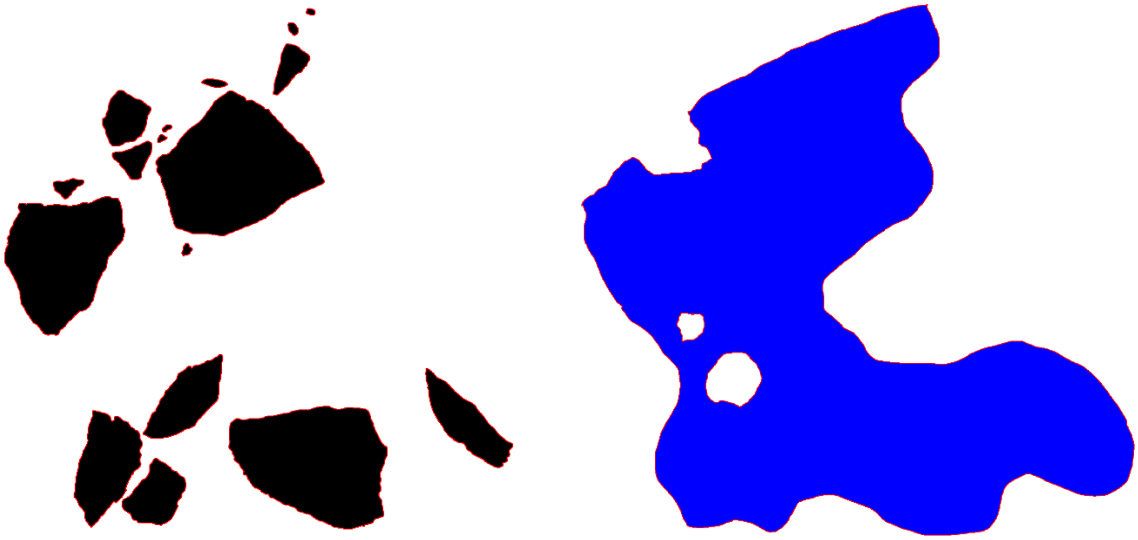
A10-240



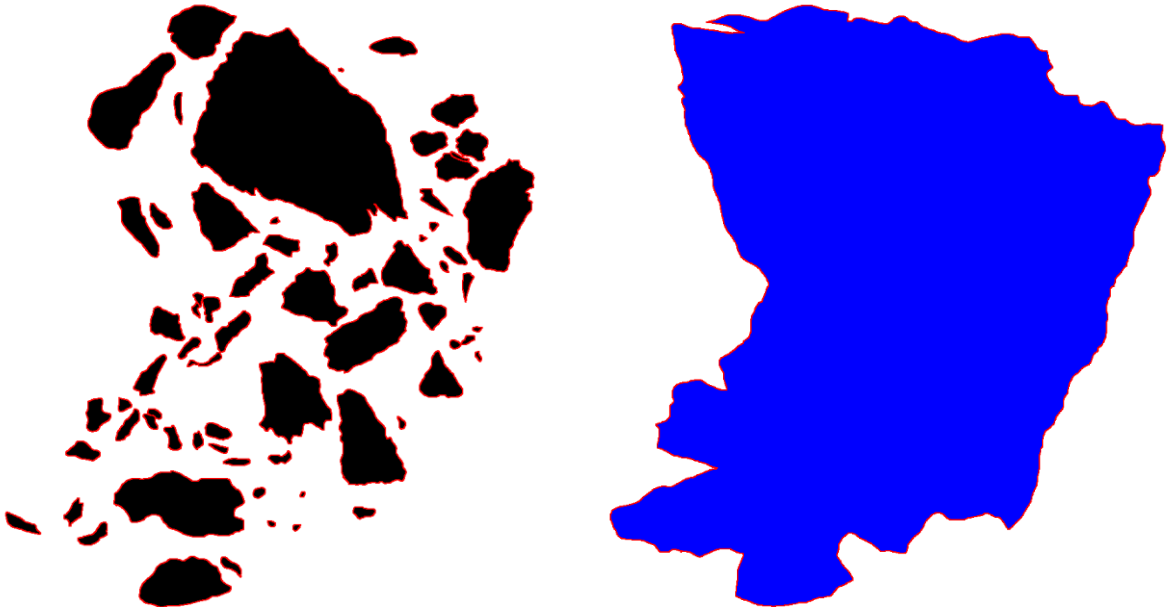
A60-240



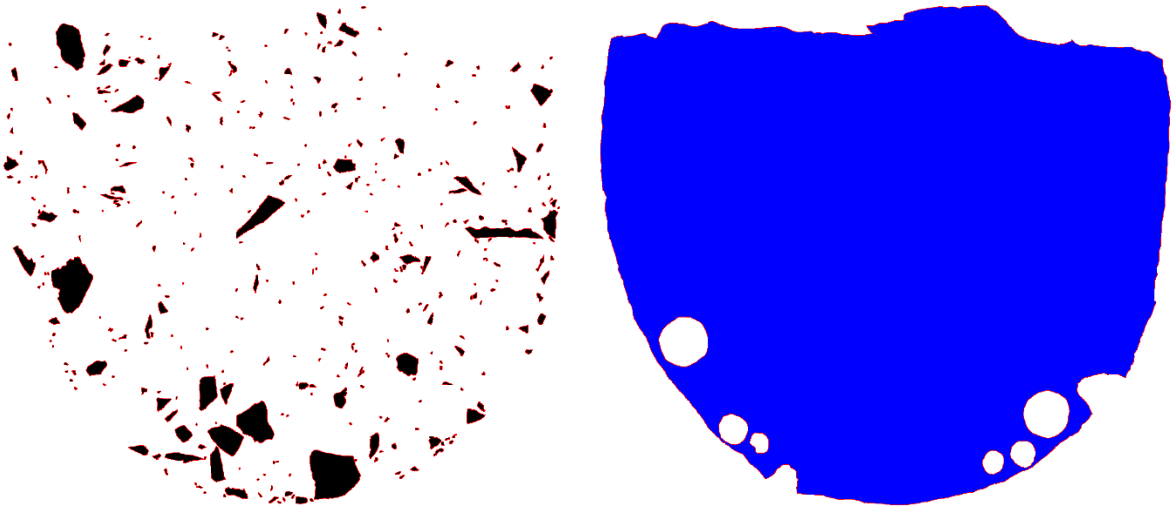
A10-480



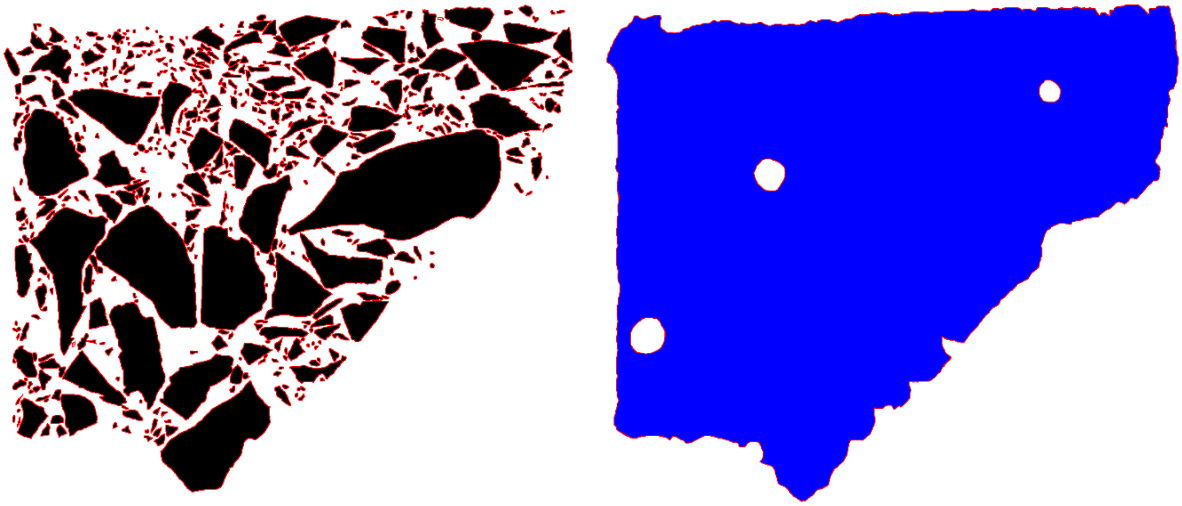
A10-1200



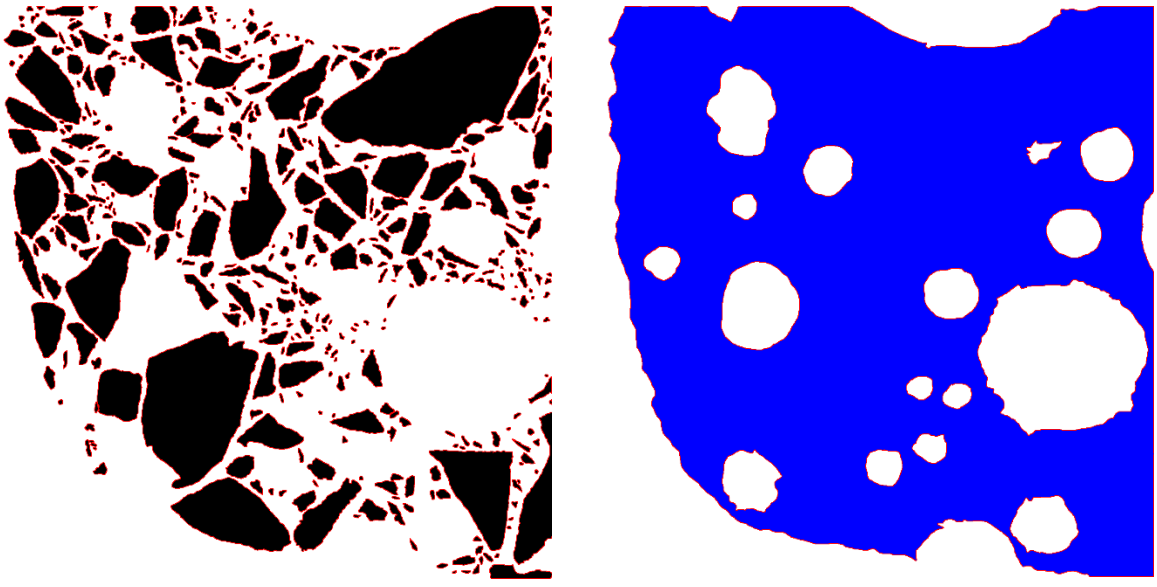
A60-1200



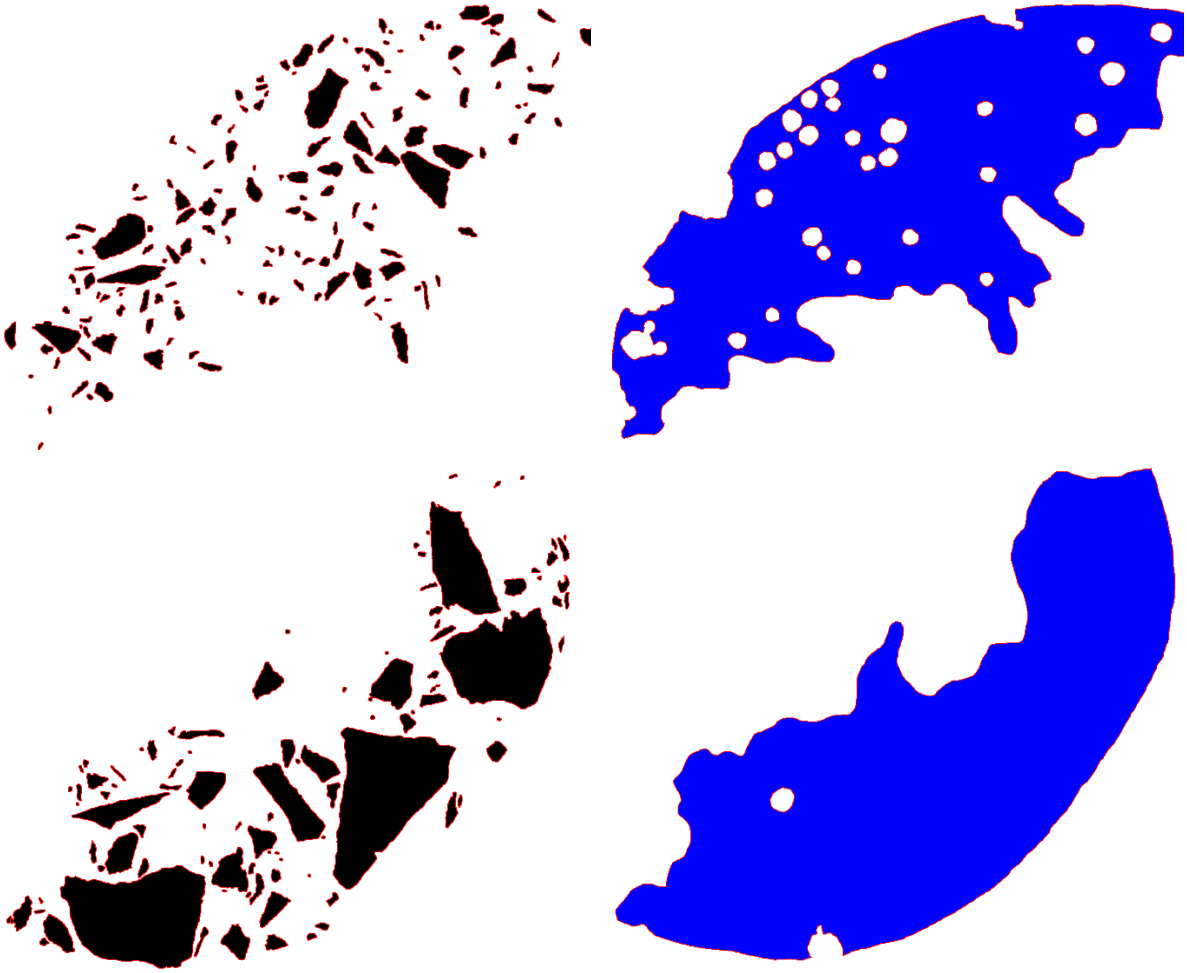
B10-22



B60-22



B60-240



B10-240 – A, B